

Encyclopedia of
Complexity and Systems Science Series
Editor-in-Chief: Robert A. Meyers

SPRINGER NATURE
Reference



Robert I. Tilling
Editor

Complexity in Tsunamis, Volcanoes, and their Hazards

A Volume in the Encyclopedia of
Complexity and Systems Science,
Second Edition

Encyclopedia of Complexity and Systems Science Series

Editor-in-Chief

Robert A. Meyers

The Encyclopedia of Complexity and Systems Science series of topical volumes provides an authoritative source for understanding and applying the concepts of complexity theory, together with the tools and measures for analyzing complex systems in all fields of science and engineering. Many phenomena at all scales in science and engineering have the characteristics of complex systems, and can be fully understood only through the transdisciplinary perspectives, theories, and tools of self-organization, synergetics, dynamical systems, turbulence, catastrophes, instabilities, nonlinearity, stochastic processes, chaos, neural networks, cellular automata, adaptive systems, genetic algorithms, and so on. Examples of near-term problems and major unknowns that can be approached through complexity and systems science include: The structure, history and future of the universe; the biological basis of consciousness; the integration of genomics, proteomics and bioinformatics as systems biology; human longevity limits; the limits of computing; sustainability of human societies and life on earth; predictability, dynamics and extent of earthquakes, hurricanes, tsunamis, and other natural disasters; the dynamics of turbulent flows; lasers or fluids in physics, microprocessor design; macromolecular assembly in chemistry and biophysics; brain functions in cognitive neuroscience; climate change; ecosystem management; traffic management; and business cycles. All these seemingly diverse kinds of phenomena and structure formation have a number of important features and underlying structures in common. These deep structural similarities can be exploited to transfer analytical methods and understanding from one field to another. This unique work will extend the influence of complexity and system science to a much wider audience than has been possible to date.

More information about this series at <https://link.springer.com/bookseries/15581>

Robert I. Tilling
Editor

Complexity in Tsunamis, Volcanoes, and their Hazards

A Volume in the Encyclopedia of Complexity
and Systems Science, Second Edition

With 354 Figures and 15 Tables

 Springer

Editor

Robert I. Tilling
Volcano Science Center
US Geological Survey
Menlo Park, USA

ISBN 978-1-0716-1704-5 ISBN 978-1-0716-1705-2 (eBook)
ISBN 978-1-0716-1706-9 (print and electronic bundle)
<https://doi.org/10.1007/978-1-0716-1705-2>

© Springer Science+Business Media, LLC, part of Springer Nature 2022

All rights are reserved by the Publisher, whether the whole or part of the material is concerned, specifically the rights of translation, reprinting, reuse of illustrations, recitation, broadcasting, reproduction on microfilms or in any other physical way, and transmission or information storage and retrieval, electronic adaptation, computer software, or by similar or dissimilar methodology now known or hereafter developed.

The use of general descriptive names, registered names, trademarks, service marks, etc. in this publication does not imply, even in the absence of a specific statement, that such names are exempt from the relevant protective laws and regulations and therefore free for general use.

The publisher, the authors, and the editors are safe to assume that the advice and information in this book are believed to be true and accurate at the date of publication. Neither the publisher nor the authors or the editors give a warranty, expressed or implied, with respect to the material contained herein or for any errors or omissions that may have been made. The publisher remains neutral with regard to jurisdictional claims in published maps and institutional affiliations.

This Springer imprint is published by the registered company Springer Science+Business Media, LLC, part of Springer Nature.

The registered company address is: 1 New York Plaza, New York, NY 10004, U.S.A.

Series Preface

The Encyclopedia of Complexity and System Science Series is a multivolume authoritative source for understanding and applying the basic tenets of complexity and systems theory as well as the tools and measures for analyzing complex systems in science, engineering, and many areas of social, financial, and business interactions. It is written for an audience of advanced university undergraduate and graduate students, professors, and professionals in a wide range of fields who must manage complexity on scales ranging from the atomic and molecular to the societal and global.

Complex systems are systems that comprise many interacting parts with the ability to generate a new quality of collective behavior through selforganization, e.g., the spontaneous formation of temporal, spatial, or functional structures. They are therefore adaptive as they evolve and may contain self-driving feedback loops. Thus, complex systems are much more than a sum of their parts. Complex systems are often characterized as having extreme sensitivity to initial conditions as well as emergent behavior that are not readily predictable or even completely deterministic. The conclusion is that a reductionist (bottom-up) approach is often an incomplete description of a phenomenon.

This recognition that the collective behavior of the whole system cannot be simply inferred from the understanding of the behavior of the individual components has led to many new concepts and sophisticated mathematical and modeling tools for application to many scientific, engineering, and societal issues that can be adequately described only in terms of complexity and complex systems.

Examples of Grand Scientific Challenges which can be approached through complexity and systems science include: the structure, history, and future of the universe; the biological basis of consciousness; the true complexity of the genetic makeup and molecular functioning of humans (genetics and epigenetics) and other life forms; human longevity limits; unification of the laws of physics; the dynamics and extent of climate change and the effects of climate change; extending the boundaries of and understanding the theoretical limits of computing; sustainability of life on the earth; workings of the interior of the earth; predictability, dynamics, and extent of earthquakes, tsunamis, and other natural disasters; dynamics of turbulent flows and the motion of granular materials; the structure of atoms as expressed in the Standard Model and the formulation of the Standard Model and gravity into a Unified Theory; the structure of water; control of global infectious diseases; and also evolution and quantification of (ultimately) human cooperative behavior in politics,

economics, business systems, and social interactions. In fact, most of these issues have identified nonlinearities and are beginning to be addressed with nonlinear techniques, e.g., human longevity limits, the Standard Model, climate change, earthquake prediction, workings of the earth's interior, natural disaster prediction, etc.

The individual complex systems mathematical and modeling tools and scientific and engineering applications that comprised the *Encyclopedia of Complexity and Systems Science* are being completely updated and the majority will be published as individual books edited by experts in each field who are eminent university faculty members.

The topics are as follows:

Agent Based Modeling and Simulation
 Applications of Physics and Mathematics to Social Science
 Cellular Automata, Mathematical Basis of
 Chaos and Complexity in Astrophysics
 Climate Modeling, Global Warming, and Weather Prediction
 Complex Networks and Graph Theory
 Complexity and Nonlinearity in Autonomous Robotics
 Complexity in Computational Chemistry
 Complexity in Earthquakes, Tsunamis, and Volcanoes, and Forecasting and
 Early Warning of Their Hazards
 Computational and Theoretical Nanoscience
 Control and Dynamical Systems
 Data Mining and Knowledge Discovery
 Ecological Complexity
 Ergodic Theory
 Finance and Econometrics
 Fractals and Multifractals
 Game Theory
 Granular Computing
 Intelligent Systems
 Nonlinear Ordinary Differential Equations and Dynamical Systems
 Nonlinear Partial Differential Equations
 Percolation
 Perturbation Theory
 Probability and Statistics in Complex Systems
 Quantum Information Science
 Social Network Analysis
 Soft Computing
 Solitons
 Statistical and Nonlinear Physics
 Synergetics
 System Dynamics
 Systems Biology

Each entry in each of the Series books was selected and peer reviews organized by one of our university-based book Editors with advice and consultation provided by our eminent Board Members and the Editor-in-Chief.

This level of coordination assures that the reader can have a level of confidence in the relevance and accuracy of the information far exceeding than that generally found on the World Wide Web. Accessibility is also a priority and for this reason each entry includes a glossary of important terms and a concise definition of the subject. In addition, we are pleased that the mathematical portions of our Encyclopedia have been selected by Math Reviews for indexing in MathSciNet. Also, ACM, the world's largest educational and scientific computing society, recognized our *Computational Complexity: Theory, Techniques, and Applications* book, which contains content taken exclusively from the *Encyclopedia of Complexity and Systems Science*, with an award as one of the notable Computer Science publications. Clearly, we have achieved prominence at a level beyond our expectations, but consistent with the high quality of the content!

Palm Desert, CA, USA
August 2021

Robert A. Meyers
Editor-in-Chief

Volume Preface

Tsunamis, volcanic eruptions, and earthquakes are natural phenomena – often interrelated – produced by the interaction of geophysical and geochemical processes operative within our dynamic Earth. Such phenomena, if sufficiently energetic and occurring in populated or cultivated regions, can be highly hazardous to people and societal infra-structure. Thus, the prerequisite for any effective program to mitigate the potential risks posed by natural hazards is the best-possible scientific understanding of the phenomena involved. Tsunamis occur when a large volume of water (in an ocean or a lake) is displaced suddenly in response to an impulsive disruption of its normal circulation regime. Most tsunamis are triggered by large plate-boundary, tectonic earthquakes that involve significant ground displacements, but they also can be generated by submarine landslides and eruption-related impacts, such as volcano flank or caldera collapse, or the rapid entry of eruptive products into water bodies. Major tsunamis can devastate coastal communities, causing many fatalities and widespread massive destruction. For example, the 2004 Indian Ocean tsunami, generated by a magnitude 9.1 earthquake, affected more than a dozen countries in the region and killed ~300,000 people. Comparatively speaking, volcanic eruptions generally are less destructive and lethal than tsunamis; the deadliest eruption in recorded history (Tambora Volcano, Indonesia, 1815) claimed ~60,000 fatalities.

Well-studied historical volcanic eruptions have shown that they are almost always preceded and accompanied by precursory activity (also called “volcano unrest”), manifested by above-background seismicity, ground deformation, and/or increased gas emission. Thus, it is essential to conduct volcano-monitoring studies – utilizing seismic, geodetic, and other geophysical or geochemical approaches – to detect the onset of volcano unrest and then to track its varying behavior with time. A key element of seismic monitoring is to characterize the nature and source of the “volcano-seismic” signals, which are distinct from those of purely tectonic, nonvolcanic origin. Since the 1970s, much has been learned about such seismic signals, and a fundamental goal of “volcano seismology” studies is to seismically characterize the configuration of active magmatic systems and to understand how they work before, during, and after eruptions. Data from volcano-monitoring studies constitute the only scientifically valid basis to make a short-term forecast of future eruption at a volcano currently in repose, or to anticipate possible changes during an ongoing eruption at an active volcano.

With the emergence of modern volcanology in the early twentieth century and the increased scientific data now available, the risk from volcanic hazards can be more precisely anticipated and assessed and be effectively mitigated by timely warnings and actions by emergency management officials. Likewise, the risk posed by tsunami hazards is now better mitigated since the establishment of several tsunami warning centers, which utilize information from seismic data, sea-level gauges, and buoy stations to generate models that more precisely forecast tsunami arrival times and estimate coastal impacts. Over recent decades, volcano-monitoring networks and tsunami warning systems have increasingly relied on satellite telecommunications, for collecting data as well as transmitting alert notifications. Moreover, advances in computer technology – both hardware and software – have greatly facilitated the acquisition and processing of the data needed for computer modeling of eruptive and tsunami processes and associated hazards.

The chapters of this book provide a selective sampling of recent studies of volcanic eruptions and tsunamis, and they clearly demonstrate that substantial strides have been made in the characterization and understanding of these phenomena, and their associated impacts and hazards. It also should be emphasized that scientific data – regardless of quality or quantity – used to declare forecasts and hazards warnings must be communicated effectively to, and acted upon in a timely manner by, emergency management authorities. As emphasized in some of the chapters, scientists still face some major challenges with regard to timely and effective communications of warnings to governmental decision makers and the affected public.

CA, USA
August 2021

Robert I. Tilling
Volume Editor

Contents

Part I Tsunami Processes, Hazards, and Forecasting	1
Tsunami Earthquakes	3
J. Polet and H. Kanamori	
Tsunamis: Stochastic Models of Occurrence and Generation Mechanisms	25
Eric L. Geist, David D. Oglesby and Kenny J. Ryan	
Wedge Mechanics: Relation with Subduction Zone Earthquakes and Tsunamis	55
Kelin Wang, Yan Hu and Jiangheng He	
Tsunamis, Inverse Problem of	71
Kenji Satake	
Tsunamis: Bayesian Probabilistic Analysis	91
Anita Grezio, Stefano Lorito, Tom Parsons and Jacopo Selva	
Tsunami Inundation, Modeling of	117
Patrick J. Lynett	
Tsunami Sedimentology	135
Pedro J. M. Costa and S. Dawson	
Tsunami from the Storegga Landslide	153
Stein Bondevik	
Tsunamis Effects in Man-Made Environment	187
Harry Yeh, Andre Barbosa and Benjamin H. Mason	
Tsunami Hazard and Risk Assessment on the Global Scale	213
F. Løvholt, J. Griffin and M. A. Salgado-Gálvez	
Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data	247
Stefano Lorito, F. Romano and T. Lay	

Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings	299
Barry Hirshorn, Stuart Weinstein, Dailin Wang, Kanoa Koyanagi, Nathan Becker and Charles McCreery	
Tsunami Forecasting and Warning	335
Osamu Kamigaichi	
Fukushima Catastrophe: The Challenge of Complexity (Collective Reflexivity, Adaptive Knowledge, Political Innovation)	373
Alain-Marc Rieu	
Part II Volcanic Processes, Eruptions, and Hazards	391
Volcano Seismology: An Introduction	393
Vyacheslav M. Zobin	
Source Quantification of Volcanic-Seismic Signals	425
Hiroyuki Kumagai	
Volcanoes, Non-linear Processes in	469
Bernard Chouet	
Volcano Deformation: Insights into Magmatic Systems	503
Daniel Dzurisin	
Volcanoes in Iceland and Crustal Deformation Processes	539
Sigrún Hreinsdóttir	
Volcanic Eruptions, Explosive: Experimental Insights	561
Stephen J. Lane and Michael R. James	
Volcanic Eruptions: Cyclicity During Lava Dome Growth	619
Oleg Melnik, R. Stephen J. Sparks, Antonio Costa and Alexei A. Barmin	
Volcanic Eruptions: Stochastic Models of Occurrence Patterns ...	647
Mark S. Bebbington	
Volcanic Hazards and Early Warning	699
Robert I. Tilling	
Volcanic Hazards Warnings: Effective Communications of	717
C. J. Fearnley	
Index	743

About the Editor-in-Chief



Robert A. Meyers

President: RAMTECH Limited

Manager: Chemical Process Technology, TRW Inc.

Postdoctoral Fellow: California Institute of Technology

Ph.D. Chemistry, University of California at Los Angeles

B.A. Chemistry, California State University, San Diego

Biography

Dr. Meyers was manager of Energy and Environmental Projects at TRW (now Northrop Grumman) in Redondo Beach, CA, and is now president of RAMTECH Limited. He is coinventor of the Gravimelt process for desulfurization and demineralization of coal for air pollution and water pollution control and was manager of the Department of Energy project leading to the construction and successful operation of a first-of-a-kind Gravimelt Process Integrated Test Plant. Dr. Meyers is the inventor of and was project manager for the DOE-sponsored Magnetohydrodynamics Seed Regeneration Project which has resulted in the construction and successful operation of a pilot plant for production of potassium formate, a chemical utilized for plasma electricity generation and air pollution control. He also managed TRW efforts in magnetohydrodynamics electricity generating combustor and plasma channel development. Dr. Meyers managed the pilot-scale DoE project for determining the hydrodynamics of synthetic fuels. He is a coinventor of several thermo-oxidative stable polymers which have achieved commercial success such as the GE PEI, Upjohn Polyimides, and Rhone-Poulenc bismaleimide resins. He has also managed projects for photochemistry, chemical lasers, flue gas scrubbing, oil shale

analysis and refining, petroleum analysis and refining, global change measurement from space satellites, analysis and mitigation (carbon dioxide and ozone), hydrometallurgical refining, soil and hazardous waste remediation, novel polymer synthesis, modeling of the economics of space transportation systems, space rigidizable structures, and chemiluminescence-based devices.

He is a senior member of the American Institute of Chemical Engineers, member of the American Physical Society, and member of the American Chemical Society and served on the UCLA Chemistry Department Advisory Board. He was a member of the joint USA-Russia working group on air pollution control and the EPA-sponsored Waste Reduction Institute for Scientists and Engineers.

Dr. Meyers has more than 20 patents and 50 technical papers in the fields of photochemistry, pollution control, inorganic reactions, organic reactions, luminescence phenomena, and polymers. He has published in primary literature journals including *Science* and the *Journal of the American Chemical Society*, and is listed in *Who's Who in America* and *Who's Who in the World*. Dr. Meyers' scientific achievements have been reviewed in feature articles in the popular press in publications such as *The New York Times Science Supplement* and *The Wall Street Journal* as well as more specialized publications such as *Chemical Engineering* and *Coal Age*. A public service film was produced by the Environmental Protection Agency on Dr. Meyers' chemical desulfurization invention for air pollution control.

Dr. Meyers is the author or editor-in-chief of a wide range of technical books including the *Handbook of Chemical Production Processes*, *Handbook of Synfuels Technology*, *Handbook of Petroleum Refining Processes* (now in fourth edition), and *Handbook of Petrochemical Production Processes* (now in second edition) (McGraw-Hill); *Handbook of Energy Technology and Economics*, published by John Wiley & Sons; *Coal Structure*, published by Academic Press; and *Coal Desulfurization* as well as the *Coal Handbook* published by Marcel Dekker. He served as chairman of the Advisory Board for *A Guide to Nuclear Power Technology*, published by John Wiley & Sons, which won the Association of American Publishers Award as the best book in technology and engineering. He also served as editor-in-chief of three editions of the Elsevier *Encyclopedia of Physical Science and Technology*. Most recently, Dr. Meyers served as editor-in-chief of the *Encyclopedia of Analytical Chemistry* as well as *Reviews in Cell Biology and Molecular Medicine* and a book series of the same name both published by John Wiley & Sons. In addition, Dr. Meyers currently serves as editor-in-chief of two SpringerNature book series, *Encyclopedia of Complexity and Systems Science* and *Encyclopedia of Sustainability Science and Technology*.

About the Volume Editor



Robert I. Tilling After receiving his Ph.D. in Geology from Yale University, Robert I. Tilling worked for the US Geological Survey (USGS) for 42 years, mostly on studies of volcanic eruptions and their associated hazards in the USA and abroad. Although “officially” retired since 2004, he remains actively involved in volcano-hazards studies as a Scientist Emeritus with the USGS Volcano Science Center in Menlo Park, California. During his career, he has authored or coauthored more than 350 geoscience papers and abstracts, including coediting the Springer volume *Monitoring and Mitigation of Volcano Hazards*. In addition to strictly technical works, he has also produced a number of so-called general-interest publications (GIPs), specifically intended for educational purposes and public outreach. For example, he was the principal co-compiler of three editions (1989, 1994, 2006) of the 80,000 copies sold *This Dynamic Planet: World Map of Volcanoes, Earthquakes, Impact Craters, and Plate Tectonics* and was the coauthor of an accompanying GIP booklet *This Dynamic Earth: The Story of Plate Tectonics*. In between research assignments, he served several USGS managerial positions, as well as being an invited consultant to some foreign countries (e.g., Colombia, Ecuador, Indonesia, Mexico, and Peru) in connection with volcano-hazards studies.

Contributors

Andre Barbosa School of Civil and Construction Engineering, Oregon State University, Corvallis, OR, USA

Alexei A. Barmin Institute of Mechanics, Moscow State University, Moscow, Russia

Mark S. Bebbington Massey University, Palmerston North, New Zealand

Nathan Becker NOAA/NWS/Pacific Tsunami Warning Center, Ewa Beach, HI, USA

Stein Bondevik Department of Environmental Sciences, Western Norway University of Applied Sciences, Sogndal, Norway

Bernard Chouet US Geological Survey, Menlo Park, USA

Antonio Costa Earth Science Department, University of Bristol, Bristol, UK

Pedro J. M. Costa Instituto D. Luiz and Departamento de Geologia da Universidade de Lisboa, Faculdade de Ciências da, Universidade de Lisboa, Lisbon, Portugal

S. Dawson Geography, School of Social Sciences, University of Dundee, Dundee, Scotland, UK

Daniel Dzurisin 1300 S.E. Cardinal Court, U.S. Geological Survey, David A. Johnston Cascades Volcano Observatory, Vancouver, WA, USA

C. J. Fearnley Department of Science and Technology Studies, University College London, London, UK

Eric L. Geist U.S. Geological Survey, Menlo Park, CA, USA

Anita Grezio Istituto Nazionale di Geofisica e Vulcanologia, Bologna, Italy

J. Griffin Geoscience Australia, Canberra, ACT, Australia

Jiangheng He Pacific Geoscience Centre, Geological Survey of Canada, Sidney, Canada

Barry Hirshorn NOAA/NWS, Ewa Beach, HI, USA

Sigrún Hreinsdóttir University of Iceland, Reykjavík, Iceland
GNS Science, Wellington, New Zealand

Yan Hu School of Earth and Ocean Sciences, University of Victoria, Victoria, Canada

Michael R. James Lancaster Environment Centre, Lancaster University, Lancaster, UK

Osamu Kamigaichi Japan Meteorological Agency, Tokyo, Japan

H. Kanamori Seismological Laboratory, Caltech, Pasadena, CA, USA

Kanoa Koyanagi NOAA/NWS/Pacific Tsunami Warning Center, Ewa Beach, HI, USA

Hiroyuki Kumagai Graduate School of Environmental Studies, Nagoya University, Nagoya, Japan

Stephen J. Lane Lancaster Environment Centre, Lancaster University, Lancaster, UK

T. Lay Department of Earth and Planetary Sciences, University of California Santa Cruz, Santa Cruz, CA, USA

Stefano Lorito Istituto Nazionale di Geofisica e Vulcanologia, Rome, Italy

F. Løvholt Norwegian Geotechnical Institute (NGI), Oslo, Norway

Patrick J. Lynett Texas A&M University, College Station, TX, USA

Benjamin H. Mason School of Civil and Construction Engineering, Oregon State University, Corvallis, OR, USA

Charles McCreery NOAA/NWS/Pacific Tsunami Warning Center, Ewa Beach, HI, USA

Oleg Melnik Institute of Mechanics, Moscow State University, Moscow, Russia

David D. Oglesby Department of Earth Sciences, University of California, Riverside, CA, USA

Tom Parsons USGS, Menlo Park, CA, USA

J. Polet Geological Sciences Department, California State Polytechnic University, Pomona, CA, USA

Alain-Marc Rieu Department of philosophy, University of Lyon-Jean Moulin, Lyon, France

Institute of East-Asian Studies (CNRS), Ecole Normale Supérieure de Lyon, Lyon, France

F. Romano Roma 1, Sez. Sismologia e Tettonofisica, Istituto Nazionale di Geofisica e Vulcanologia, Roma, Italy

Kenny J. Ryan Air Force Research Laboratory, Albuquerque, NM, USA

M. A. Salgado-Gálvez Centre Internacional de Metodes Numerics en Enginyeria (CIMNE), Barcelona, Spain

Kenji Satake Earthquake Research Institute, The University of Tokyo, Tokyo, Japan

Jacopo Selva Istituto Nazionale di Geofisica e Vulcanologia, Bologna, Italy

R. Stephen J. Sparks Earth Science Department, University of Bristol, Bristol, UK

Robert I. Tilling Volcano Science Center, US Geological Survey, Menlo Park, USA

Dailin Wang NOAA/NWS/Pacific Tsunami Warning Center, Ewa Beach, HI, USA

Kelin Wang Pacific Geoscience Centre, Geological Survey of Canada, Sidney, Canada

School of Earth and Ocean Sciences, University of Victoria, Victoria, Canada

Stuart Weinstein NOAA/NWS/Pacific Tsunami Warning Center, Ewa Beach, HI, USA

Harry Yeh School of Civil and Construction Engineering, Oregon State University, Corvallis, OR, USA

Vyacheslav M. Zobin Observatorio Vulcanológico, Universidad de Colima, Colima, Mexico

Part I

Tsunami Processes, Hazards, and Forecasting

Tsunami Earthquakes

J. Polet¹ and H. Kanamori²

¹Geological Sciences Department, California State Polytechnic University, Pomona, CA, USA

²Seismological Laboratory, Caltech, Pasadena, CA, USA

Article Outline

Glossary

Definition of the Subject and Its Importance

Introduction

Characteristics of Tsunami Earthquakes

Factors Involved in the Seismogenesis and Tsunamigenesis of Tsunami Earthquakes

A Model for Tsunami Earthquakes

Future Directions

Bibliography

Glossary

Magnitude saturation Due to the shape of the seismic source spectrum, relatively short-period measurements of seismic magnitude will produce similar magnitudes for all earthquakes above a certain size. The value of this threshold earthquake size depends on the period of the measurement: magnitude measurements using shorter-period waves will saturate at lower values than magnitude measurements using longer-period waves. M_w will not saturate.

m_b Body wave magnitude, based on the amplitude of the direct P-wave, period of the measurement: 1.0–5.0 s. Also see: Seismic magnitude.

M_S Surface wave magnitude, based on the amplitude of surface waves, period of the measurement: 20 s. Also see: Seismic magnitude.

M_w Moment magnitude, determined from the seismic moment of an earthquake, typical period of the measurement: >200 s. Also see: Seismic magnitude.

Run-up height Difference between the elevation of maximum tsunami penetration (inundation line) and the sea level at the time of the tsunami.

Seismic magnitude A scale for the relative size of earthquakes. Many different scales have been developed, almost all based on the logarithmic amplitude of a particular seismic wave on a particular type of seismometer, with corrections for the distance between source and receiver. These measurements are made for different wave types at different frequencies and thus may lead to different values for magnitude for any one earthquake.

Seismic moment The product of the fault surface area of the earthquake, the rigidity of the rock surrounding the fault, and the average slip on the fault.

Tsunami earthquake An earthquake that directly causes a regional and/or teleseismic tsunami that is greater in amplitude than would be expected from its seismic moment magnitude.

Tsunami magnitude A scale for the relative size of tsunamis generated by different earthquakes, M_t in particular is calculated from the logarithm of the maximum amplitude of the tsunami wave measured by a tide gauge distant from the tsunami source, corrected for the distance to the source (also see Satake, this volume, ► “Tsunamis, Inverse Problem of”).

Definition of the Subject and Its Importance

The original definition of “tsunami earthquake” was given by Kanamori (1972) as “an earthquake that produces a large size tsunami relative to the value of its surface wave magnitude (M_S).” Therefore, the true damage potential that a tsunami earthquake represents may not be recognized by conventional near real-time seismic analysis methods and may only become apparent upon the arrival of the tsunami waves on the local coastline. Although tsunami earthquakes occur relatively infrequently, the effect on the local population

can be devastating, as was recently illustrated by the 2010 M_w 7.8 Mentawai earthquake. This event struck offshore western Indonesia, created a tsunami with 5–9 m local run-up, and caused more than 400 casualties (Newman et al. 2011).

It is important to note that the definition of “tsunami earthquake” is distinct from that of “tsunamigenic earthquake.” A tsunamigenic earthquake is any earthquake that excites a tsunami. Tsunami earthquakes are a specific subset of tsunamigenic earthquakes, which we will later in this article more precisely define as earthquakes that directly cause a regional and/or teleseismic tsunami that is greater in amplitude than would be expected from their seismic moment magnitude.

Introduction

Shallow oceanic earthquakes may excite destructive tsunamis. Truly devastating tsunamis occur only infrequently but, as the natural disasters of the tsunamis following the 2004 Sumatra-Andaman Islands and 2011 Tohoku earthquakes have shown, may cause widespread damage and lead to tens to hundreds of thousands of casualties. In general, tsunamis are caused by shallow earthquakes beneath the ocean floor displacing large volumes of water. Thus, the magnitude of the earthquake plays an important role in determining its tsunamigenic potential. However, a particular subclass of shallow subduction zone earthquakes, “tsunami earthquakes,” poses a special problem.

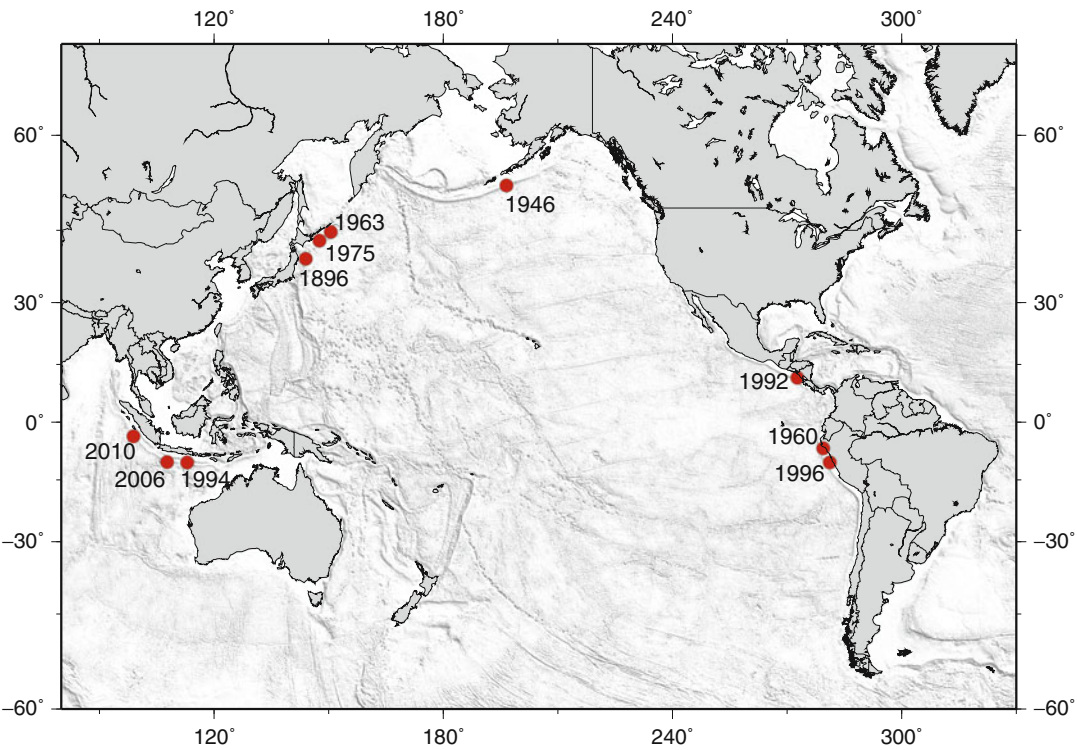
For the purpose of this article, we will define the term “tsunami earthquake” as follows: an earthquake that directly causes a regional and/or teleseismic tsunami that is greater in amplitude than would be expected from its seismic moment magnitude. With this definition we specifically exclude seismic events that were followed by tsunamis directly caused by slides or slumps resulting from the original earthquake (e.g., as was the case for the 1992 Flores (Hidayat et al. 1995; Imamura et al. 1995) and the 1998 Papua New Guinea earthquakes (Geist 2000; Synolakis et al. 2002)). We further exclude events that only very locally caused large tsunamis as a result of, for example, focusing effects due to features of the ocean floor

bathymetry (e.g., Satake and Kanamori 1991) or directivity effects combined with the shape of the coastline, as was the case for the tsunamis that hit the harbor of Crescent City after the 1964 Alaska earthquake (Brown 1964) and the November 15, 2006, Kurile Island event (Dengler et al. 2009). Furthermore, this definition compares the size of the tsunami with the moment magnitude of the earthquake and not its surface wave magnitude, slightly modifying the definition given by Kanamori (1972), in order to exclude great events for which the surface wave magnitude saturates.

Our primary objective in this entry is to describe the characteristics of tsunami earthquakes and the possible factors involved in the anomalously strong excitation of tsunamis by these events. We will also discuss possible models for these infrequent, but potentially very hazardous, events. The earthquakes listed in Table 1 and plotted in Fig. 1 are considered tsunami earthquakes, according to our modified definition presented in the previous paragraph, by the majority of the community of earthquake and tsunami researchers. However, we note that the interpretation of the 1994 Java and the 1946 Aleutian Islands earthquakes varies with investigators. The 1994 Java earthquake occurred off the southeastern coast of this island, near the east end of the Java Trench in the Indian Ocean, at 1:18 am local time. It generated a devastating tsunami that took the lives of more than 200 East Java coastal residents. Run-up measured along the southeastern Java coast ranged from 1 to 14 m, while run-up measured along the southwestern coast of Bali ranged from 1 to 5 m (Synolakis et al. 1995; Tsuji et al. 1995). Although the anomalously high tsunami excitation of the 1994 event is not in doubt, its earthquake source characteristics have been debated (Polet and Kanamori 2000; Abercrombie et al. 2001). The 1946 Aleutian Islands earthquake off Unimak Island produced one of the largest trans-Pacific tsunamis and had a tsunami magnitude of 9.3 (Abe 1979), but its moment magnitude is only $M_w = 8.2$, making it a tsunami earthquake (Johnson and Satake 1997). Some of the great tsunami heights measured (exceeding 30 m in height on Unimak Island and 16 m in run-up at the Hawaiian Islands (Tanioka and Seno 2001b)) can be attributed to slumping (Fryer et al. 2004).

Tsunami Earthquakes, Table 1 Tsunami earthquakes (references for most entries are listed in main text; others are from the National Geophysical Data Center Global Tsunami Database (http://www.ngdc.noaa.gov/hazard/tsu_db.shtml) and the Centennial Earthquake Catalog (Engdahl and Villaseñor 2002))

Date	Geographical region	M _w	m _b	M _S	M _t	Deaths
1896/06/15	Japan			7.2	8.0	26,360
1946/04/01	Aleutian Islands	8.2		7.3	9.3	165
1960/11/20	Peru	7.6	7.0	7.0		66
1963/10/20	Kurile Islands	7.8	7.1	7.2		
1975/06/10	Kurile Islands	7.5	5.6	7.0		
1992/09/02	Nicaragua	7.7	5.4	7.2		179
1994/06/02	Java	7.8	5.7	7.1		250
1996/02/21	Peru	7.5	5.8	6.6		12
2006/07/17	Java	7.7	6.2	7.2		802
2010/10/25	Mentawai, Indonesia	7.8	6.5	7.3		431



Tsunami Earthquakes, Fig. 1 Map of tsunami earthquakes (listed in Table 1). Location for 1896 earthquake from Abe (1989) and for 2006 and 2010 earthquakes from

the Global CMT catalog. All other earthquake locations from the Centennial Earthquake Catalog (Engdahl and Villaseñor 2002)

However, its anomalously high tsunamis are probably primarily due to the seismic source directly (Lopez and Okal 2006).

Several other earthquakes in the past decade have produced damaging tsunamis and have been mentioned as possible tsunami earthquakes. Seno

and Hirata (2007) suggest that the great 2004 Sumatra-Andaman earthquake also likely involved a component of tsunami earthquakes, because tsunamis larger than expected from seismic slip occurred, possibly due to slow slip in the shallow subduction boundary. Similarly, the 2011 Tohoku earthquake involved a coseismic rupture that extended all the way to the trench (e.g., Lay et al. 2012; Satake et al. 2013) and could be considered a combination of a tsunami earthquake with an earthquake that produced deeper slip, such as the 869 Jogan earthquake in the same region (Satake et al. 2013). It has also been proposed that the Kurile Islands earthquake of November 15, 2006 (Ji et al. 2007), may have exhibited some characteristics of tsunami earthquakes. The Solomon Island earthquake of April 1, 2007, excited large tsunamis, at least locally (Chen et al. 2009). However, the disparity between seismic and tsunami excitation by these events is not nearly as large as for the events in Table 1, and we do not list these events as tsunami earthquakes.

The other, less controversial, tsunami earthquakes listed in our table are the 1896 Sanriku event near the coast of Japan; two events near the Kurile Islands, one in 1963 and the other in 1975; the 1992 Nicaragua earthquake; the Peru earthquake 4 years later, as well as an earlier event in this region in 1960; the 2006 Java earthquake; and most recently the 2010 Mentawai earthquake. The June 15, 1896, Sanriku earthquake generated devastating tsunamis with a maximum run-up of 25 m and caused the worst tsunami disaster in the history of Japan with over 20,000 deaths, despite its moderate surface wave magnitude ($M_S = 7.2$) and weak seismic intensity (Hatori 1967; Abe 1989; Scholz and Small 1997). The November 20, 1960, Peru earthquake excited a tsunami that was anomalously large for an earthquake of moderate magnitude (Pelayo and Wiens 1990), resulting in 66 fatalities (from the tsunami event database of the National Geophysical Data Center, <http://www.ngdc.noaa.gov/nndc/struts/form?t=101650&s=70&d=7>). The October 20, 1963, Kurile earthquake was an aftershock to the great Kurile Islands

underthrusting earthquake ($M_w = 8.5$) of October 13, 1963, and produced a maximum run-up height of 10–15 m at Urup Island, much larger than the height of the main shock tsunami of 5 m (Abe 1989). The 1975 earthquake occurred south of the Kurile Islands and was weakly felt along the entire southern part of the Kurile Islands. Like the 1963 tsunami earthquake, this event can be considered an aftershock (Fukao 1979) of a larger event ($M_S = 7.7$) that occurred essentially at the same location on June 17, 1973. The maximum run-up height was 5 m on Shikotan Island, while the main shock had a run-up height measured at 4.5 m. A fairly strong tsunami was also recorded on tide gauges in Alaska and Hawaii (from the tsunami event database of the National Geophysical Data Center <http://www.ngdc.noaa.gov/nndc/struts/form?t=101650&s=70&d=7>). After a time period of almost two decades without a significant tsunami earthquake, the 1992 Nicaragua earthquake was the first tsunami earthquake to be captured by modern broadband seismic networks. This tsunami caused 179 deaths (from the Emergency Disasters Database, <http://www.em-dat.net/disasters/list.php>) and significant damage to the coastal areas of Nicaragua and Costa Rica, reaching heights of up to 8 m (Satake et al. 1993). The 1996 Peru earthquake struck at 7:51 am local time, approximately 130 km off the northern coastal region of Peru. Approximately 1 h after the main shock, a damaging tsunami reached the Peruvian coast, with run-up heights of 1–5 m along a coastline of 400 km (Heinrich et al. 1998), resulting in 12 deaths (Bourgeois et al. 1999). The 2006 Java earthquake was located only about 600 km west-northwest of the tsunami earthquake that occurred 12 years earlier in the same subduction zone. More than 600 people were killed by a tsunami that had maximum run-up heights exceeding 20 m along the coast of central Java (Fritz et al. 2007; Lavigne et al. 2007). The $M_w 7.8$ earthquake that struck offshore the Mentawai Islands in western Indonesia on October 25, 2010, created a tsunami with 5–9 m local run-up that caused more than 400 casualties (Newman et al. 2011).

Characteristics of Tsunami Earthquakes

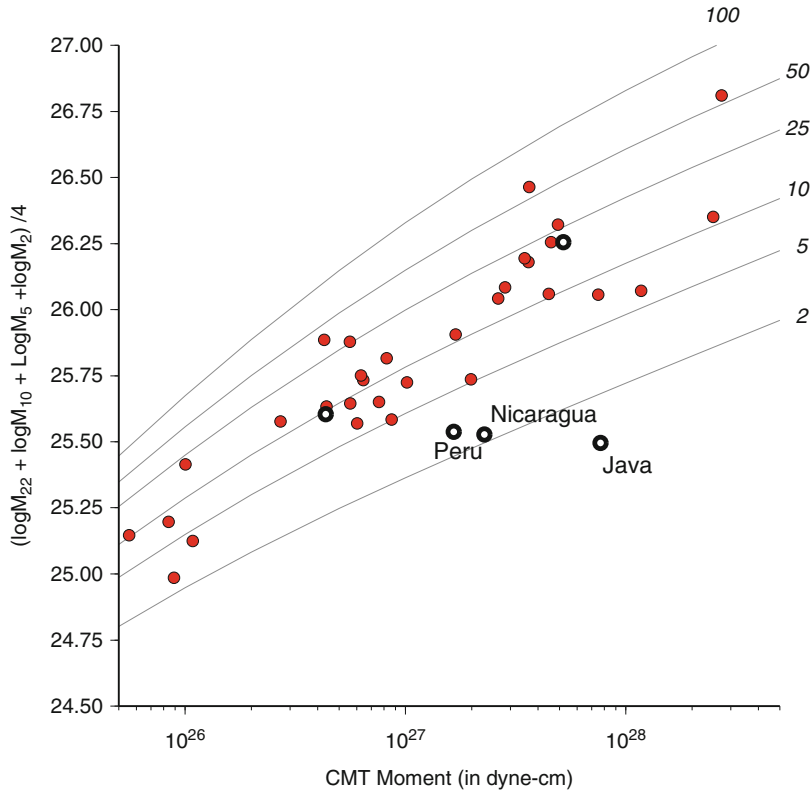
Fortunately, tsunami earthquakes occur only infrequently. Fewer than ten of these events have occurred in the last four decades since the global installation of seismic broadband instruments and tide gauges and easy availability of their data were established. However, from the detailed investigations of the most recent events and comparisons with the limited data available for the older earthquakes in Table 1, several characteristics of these earthquakes clearly emerge.

Slow Character

The slow character of tsunami earthquakes manifests itself in several different, yet related, ways. One well-established characteristic of tsunami earthquakes is the discrepancy between the determined values of the different seismic magnitude types, calculated from various kinds of seismic waves or waves of different frequency ranges. A typical comparison is that of the body wave magnitude, m_b , or the surface wave magnitude, M_S , with the moment magnitude of the earthquake, M_w . For tsunami earthquakes m_b is typically much smaller than the other two magnitudes, M_S and M_w , and M_w typically exceeds M_S . The discrepancy between these different magnitudes is more pronounced than for regular subduction zone earthquakes with similar moment magnitudes. For example, for the 1992 Nicaragua earthquake, $m_b = 5.4$, $M_S = 7.2$, and $M_w = 7.7$ (Engdahl and Villaseñor 2002); for the 1994 Java tsunami earthquake, $m_b = 5.7$, $M_S = 7.1$, and $M_w = 7.8$ (Polet and Kanamori 2000); for the 2006 Java earthquake, $m_b = 6.2$, $M_S = 7.2$, and $M_w = 7.7$ (Ammon et al. 2006); and for the 2010 Mentawai earthquake, $m_b = 6.5$, $M_S = 7.3$, and $M_w = 7.8$ (Lay et al. 2011). Since the body wave magnitude is calculated from short-period P-waves, the surface wave magnitude is determined by the amplitude of surface waves with a period of 20 s, and the moment magnitude is generally based on longer periods for big events; this consistent discrepancy is an indication of the relatively greater seismic energy release at longer periods (or the “slow” character)

of these tsunami events. Similarly, investigations of teleseismic P-waves (Polet and Kanamori 2000; Ammon et al. 2006) have shown that their source spectra are depleted in high-frequency energy at periods shorter than 20 s as compared to other shallow subduction zone earthquakes (see Fig. 2, from Polet and Kanamori 2000) as well as their own aftershocks (see Fig. 3, adapted from Polet and Thio 2003). Modeling of the rupture processes shows that the rupture velocities for tsunami earthquakes are slower than for most other subduction zone earthquakes (for several events, Pelayo and Wiens 1992; Ji 2006b; for Aleutians 1946, Lopez and Okal 2006; for Nicaragua 1992, Kanamori and Kikuchi 1993; Ide et al. 1993; for Peru 1996, Ihmlé et al. 1998; for Java 2006, Ammon et al. 2006; for Mentawai 2010, Newman et al. 2011; Lay et al. 2011). Correspondingly, the centroid times and source durations or rise times determined for these events are also relatively large with respect to other large subduction zone earthquakes (Polet and Kanamori 2000; for the Kurile Islands 1975 earthquake, Shimazaki and Geller 1977; for Peru 1960, Pelayo and Wiens 1990; for Java 2006, Ammon et al. 2006; Hara 2006; for Mentawai 2010, Newman et al. 2011; Lay et al. 2011), although they may not be anomalous relative to other, smaller, subduction zone earthquakes at very shallow depth (Bilek and Lay 2002). The energy that is radiated by these slow rupture processes is also anomalously low, as is shown by analyses of the radiated energy-to-moment ratio (for several recent tsunami earthquakes, Newman and Okal 1998; for the 1946 Aleutian Islands earthquake, Lopez and Okal 2006) and radiation efficiency (Venkataraman and Kanamori 2004).

Unfortunately, the slow character of tsunami earthquakes also means that local residents are not warned by strong ground shaking of the possibility of an impending tsunami. Field surveys and first-person accounts describe the motion of tsunami earthquakes more as a weak “rolling motion” than the usual impulsive character of local events. In the case of the Nicaragua earthquake, some felt a very feeble shock before the tsunami, but most did not feel the



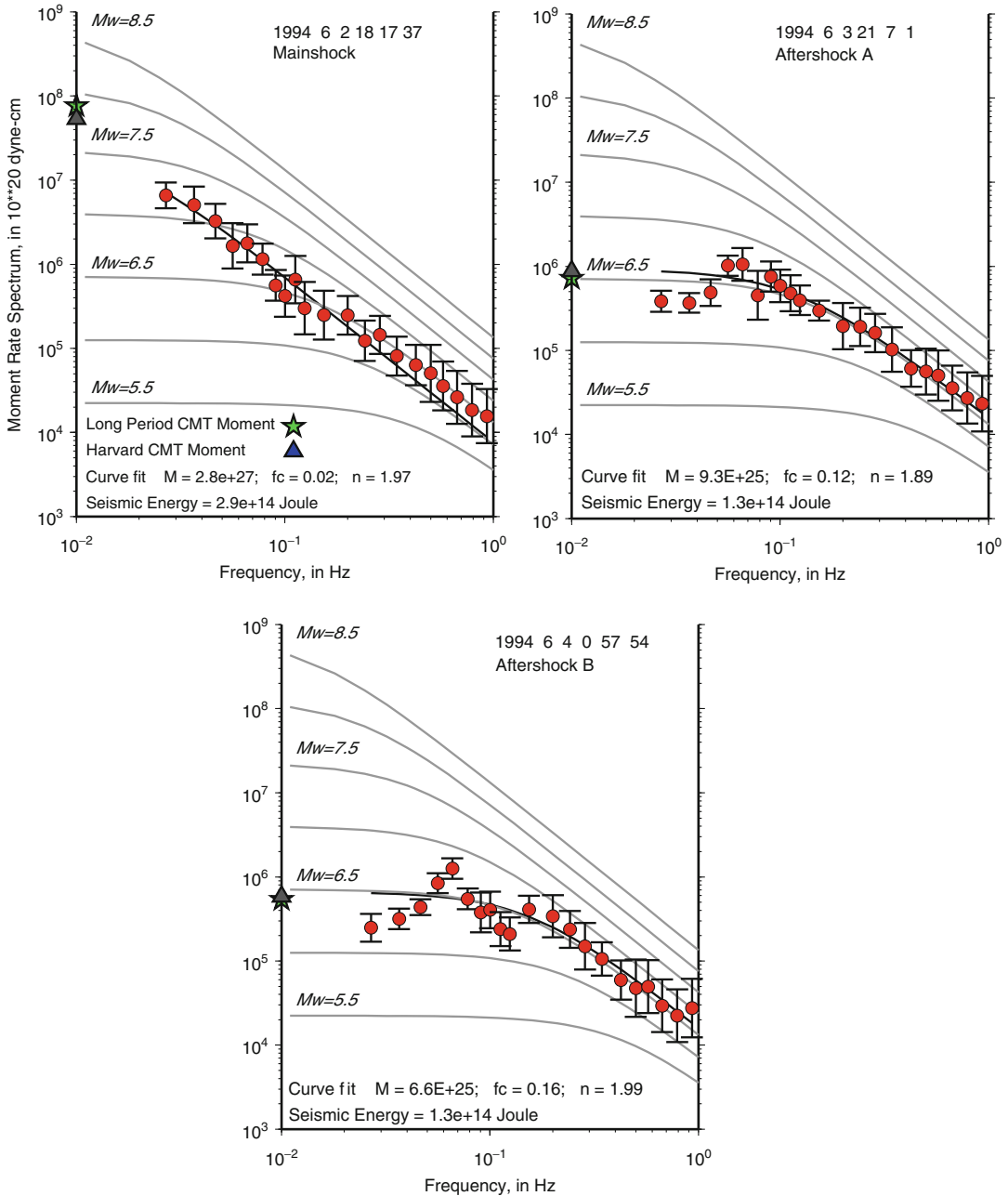
Tsunami Earthquakes, Fig. 2 Average of log of moment rate spectrum at four periods (2, 5, 10, 22 s) as a function of the seismic moment as determined by CMT inversion of long-period surface waves. Reference curves were calculated for an ω^2 model (Brune 1970). Values next to the gray curves indicate the stress parameter used to calculate the reference curve. Events shown are all shallow subduction zone earthquakes from 1992 to 1996

with moment magnitude 7.0 or greater. Earthquakes followed by anomalously large tsunamis are indicated with open circles. Of these events, only the 1992 Nicaragua, 1994 Java, and 1996 Peru earthquakes are slow tsunami earthquakes, as is shown in this figure by their relatively low moment rate spectrum at shorter periods (Adapted from Polet and Kanamori (2000))

earthquake at all (Ide et al. 1993). For the 1994 Java event, earthquake-induced ground shaking was not noticed by the coastal residents interviewed in Bali and Java (Synolakis et al. 1995). Interviews with local residents carried out for the 2006 Java earthquake (Mori et al. 2007) also indicate that they felt little or no shaking. Similarly, based on information gathered from eyewitness interviews conducted during a field survey of the effects of the 2010 Mentawai event, the earthquake was not strongly felt (Hill et al. 2012).

Most designs for tsunami earthquake discriminators and early warning systems make use of a number of the manifestations of the unusually slow character of tsunami

earthquakes listed above, and many incorporate the use of long-period seismic waves for robust estimation of the size of the event. For example, the pulse width of the P-wave, used to calculate moment magnitude M_{wp} , can give an accurate estimate of source duration time. The combination of M_{wp} and the source duration can provide an effective tool to issue early tsunami warnings (Tsuboi 2000). A slightly later arrival on a seismogram, the W-phase, is a distinct ramp-like long-period (up to 1,000 s) phase that begins between P- and S-waves on displacement seismograms and is particularly pronounced for slow earthquakes; thus it can be used for identification of these types of events (Kanamori 1993). Another method for fast



Tsunami Earthquakes, Fig. 3 Moment rate spectra for the Java thrust main shock (*left*) and two of its largest aftershocks, tensional events in the outer rise (*right panels*). The stars indicate the Harvard CMT moment, the triangles the moment determined using very long-period surface waves. Gray reference curves were

calculated for an ω^2 model (Brune 1970) with a stress parameter of 30 bars and an S-wave velocity of 3.75 km/s. The moment rate for the main shock is similar to that of its much smaller (in terms of M_w) aftershocks for periods shorter than 10 s (Adapted from Polet and Thio (2003))

regional tsunami warning uses the ratio of the total seismic energy to the high-frequency seismic energy (between 1 and 5 Hz), computed from

the seismograms (Shapiro et al. 1998). Similarly, the detection of deficient values of seismic energy-to-seismic moment ratio can be

accomplished in automated, real-time mode (Newman and Okal 1998).

Location: Close to Trench

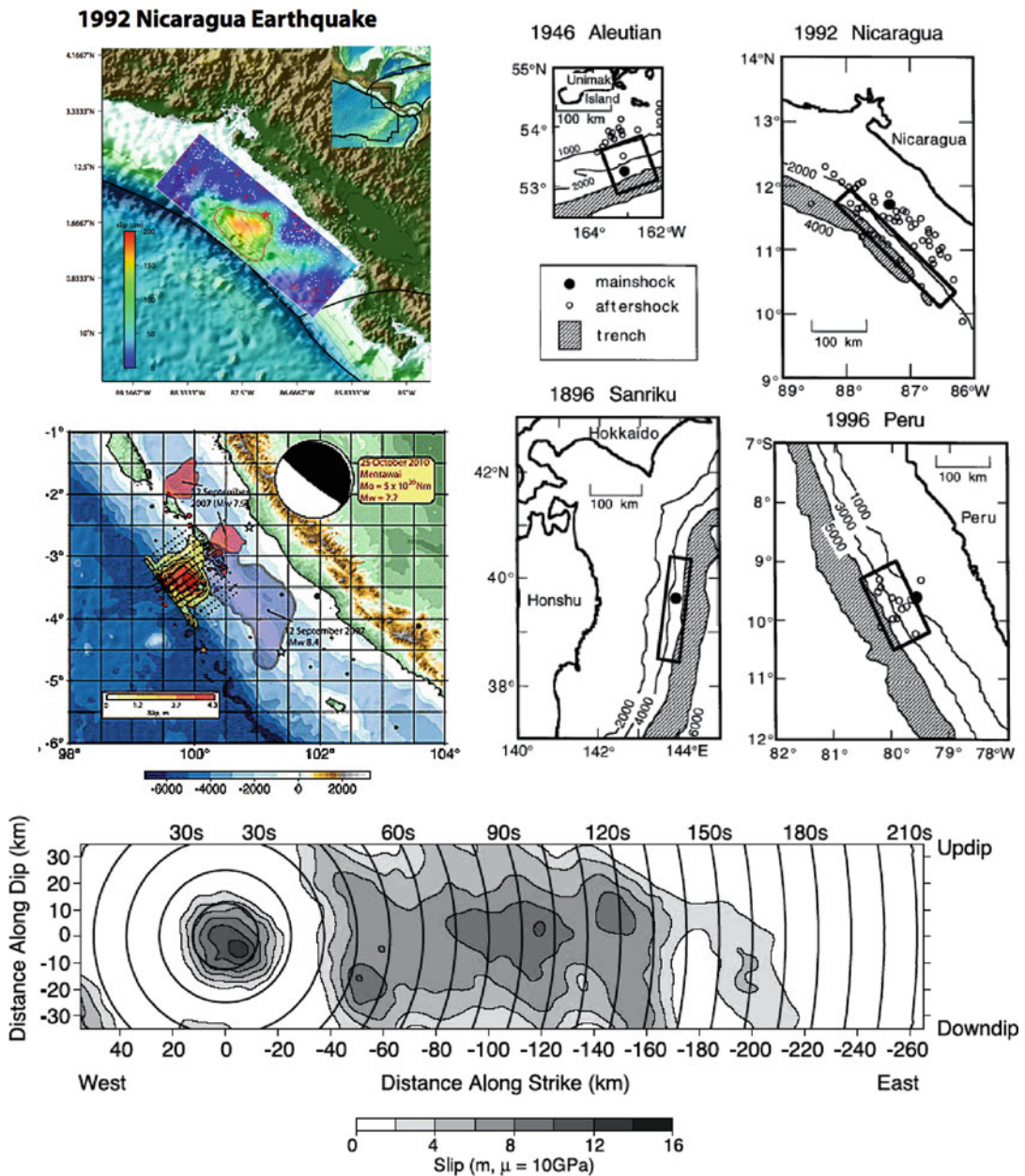
The hypocenters of the recent tsunami earthquakes are located relatively close to the trench, was compared to regular subduction zone earthquakes. The Global CMT and other (Polet and Kanamori 2000) centroid locations for several of these events are located even on the seaward side of the trench (also see Fig. 5). It may be possible that the inversion process mislocates the centroid of the event due to the unusually long duration of the seismic source for its moment and thus its unusually late centroid time. Inversions using seismic and/or tsunami waveforms and other waveform investigations for the 1896 Sanriku event (Tanioka and Satake 1996a), the 1946 Aleutian Islands earthquake (Johnson and Satake 1997; Tanioka and Seno 2001b), the 1960 Peru earthquake (Pelayo and Wiens 1992), the 1963 Kurile earthquake (Beck and Ruff 1987; Wiens 1989; Pelayo and Wiens 1992), the 1975 Kurile earthquake (Wiens 1989; Pelayo and Wiens 1992), the 2006 Java earthquake (Fujii et al. 2006; Ammon et al. 2006), the Nicaragua earthquake (Kanamori and Kikuchi 1993; Satake 1994; Kikuchi and Kanamori 1995), the 1996 Peru event (Ihmlé et al. 1998; Satake and Tanioka 1999), and the 2010 Mentawai earthquake (Hill et al. 2012; Lay et al. 2011; Newman et al. 2011) indicate the presence of concentrated slip in a narrow region near the trench (see Fig. 4). Although in most of these inversions only 1-D Green's functions were used, initial research results using a more realistic velocity model for the shallow subduction zone (Okamoto and Takenaka 2009) shows a similar picture for the 2006 Java tsunami earthquake.

The recent development of body wave back-projection techniques has led to the examination of earthquake rupture processes in a frequency-dependent manner. Investigations of the 2011 Tohoku earthquake using this approach have found that the down-dip region of the megathrust radiated higher relative levels of short-period radiation than the up-dip regime for this great earthquake (Koper et al. 2011; Simons et al. 2011).

Similar investigations of the February 27, 2010, Chile and 2004 Sumatra-Andaman great earthquakes also image sources of coherent short-period radiation in the down-dip portions of the megathrusts, with large slip up-dip of the short-period sources (see references in Lay et al. 2012). Therefore, the frequency of the seismic radiation during great subduction earthquakes appears to be related to the location of the source of this radiation, with the shallower depths of the megathrust producing relatively little coherent short-period radiation (Lay et al. 2012).

Aftershocks

The aftershock sequences of (recent) tsunami earthquakes are unusual in the preponderance of events not located on the interface between overriding and subducting plates (Polet and Kanamori 2000). Some of these aftershocks are located in the outer rise according to their Global CMT centroid locations, and, in the case of the Java 2006 aftershocks, relocations using a 2.5-D model of the subduction zone (Okamoto and Takenaka 2009) confirm this location. Others are located in the overriding plate (Bilek et al. 2007 for the 2006 Java earthquake), with some deeper within the accretionary prism (for the Java 2006 earthquake: Okamoto and Takenaka 2009). The low number, or nonexistence, of large (greater than magnitude 5.5) interplate earthquakes suggests that the main shock almost completely relieved the stress on the interface or may be related to the frictional properties of the fault. Several explanations have been proposed for the anomalously high number of intraplate earthquakes following tsunami earthquakes. Because of the proximity of the areas of high slip to the trench, the stress change in the outer rise and trench area due to a tsunami earthquake are greater than for the "standard" subduction zone earthquake. Several modeling studies of stress alterations caused by large subduction earthquakes suggest that the subduction slip will act to increase the tensional stress and favor normal events in zones toward the ocean from the upper limit of the rupture (Taylor et al. 1996; Dmowska et al. 1996). The subducting plate also seems to have a highly broken-up or rough character in



Tsunami Earthquakes, Fig. 4 Slip or fault models determined for various tsunami earthquakes. The models shown are for Nicaragua 1992 (*top left*, from Ji 2006), Mentawai 2010 (*center left*, from Lay et al. 2011), Java 2006 (*bottom*, from Ammon et al. 2006), and four tsunami

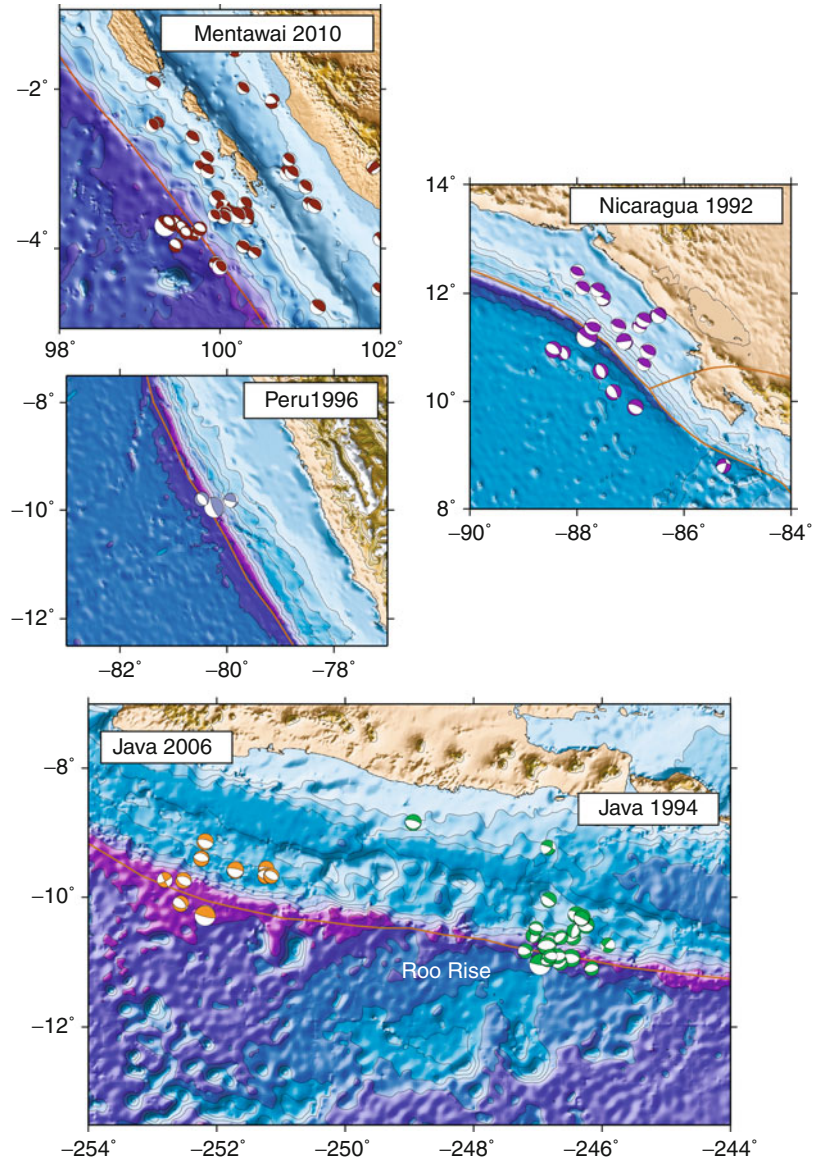
earthquakes (*top right*, from Satake and Tanioka 1999); the main shock and aftershock epicenters are shown and the hatched area indicates the trench. These models all show the presence of substantial slip close to the trench.

many areas in which tsunami earthquakes have occurred (Tanioka et al. 1997; McAdoo et al. 2004; for Java, Masson et al. 1990; for Peru,

Hilde 1983; Kulm et al. 1981), which suggests the presence of more pervasive preexisting weak zones, due to, for example, seafloor spreading

Tsunami Earthquakes,

Fig. 5 Aftershock sequences (seismicity in the region of the main shock within a period of 4 years after its occurrence) for several tsunami earthquakes, from the Global CMT catalog. Relatively few interplate aftershocks occur on the megathrust after a tsunami earthquake occurs, but a preponderance of normal fault, probably intraplate, aftershocks is apparent



related fabric. These weak zones may be reactivated in outer rise, or deeper intraplate, earthquakes following a tsunami earthquake.

Factors Involved in the Seismogenesis and Tsunamigenesis of Tsunami Earthquakes

Based on the consistent characteristics of tsunami earthquakes, as described in the previous section,

and observations of their tectonic environments, hypotheses have been developed as to the cause of their extraordinary tsunami excitation and unusual seismic source process. In this section, we will document the factors that have been proposed to affect the seismo- and tsunamigenesis of tsunami earthquakes. Some are associated with the numerical prediction of tsunami wave heights based on observed seismic waveforms and others with possible unusual conditions of the tectonic environment in which

these events occur; many of these factors are closely or at least somewhat related.

Slow Character May Lead to Underestimation of Earthquake Size

Prior to the installation of the broadband Global Seismic Network, the magnitudes of earthquakes were often determined from the amplitude of their teleseismic P-waves only. In the case of tsunami earthquakes, using this technique to determine their magnitude would lead to an initial underestimation of their true size, since their source spectra are depleted in the relatively high-frequency energy that usually dominates the direct P-wave signals of regular earthquakes (Polet and Kanamori 2000). A similar issue would occur, although to a lesser degree, when using surface waves of periods of 20 s to determine the surface wave magnitude of these slow events (Pelayo and Wiens 1992). With the advent of broadband sensors in the past several decades, it has now however become possible to investigate seismic waves to very long periods, hundreds or even thousands of seconds. Using more sophisticated techniques and the waveforms from technologically advanced sensors, the source spectrum of the recent tsunami earthquakes can now be modeled to these very long periods. Thus, no long-period seismic energy that would excite tsunami waves should be “hidden” from the view of seismologists in the computation of moment magnitude or full rupture models using long-period surface and body waves. However, for the earthquakes discussed in this entry, the observed tsunami is still larger than would be expected, even for their moment magnitude.

Effect of the Presence of Weak Materials with Low Shear Modulus

Most earthquake source inversions (either for full rupture or centroid moment tensor parameters) implement a simple one-dimensional velocity and rigidity model to compute synthetic seismograms and model the recorded waveforms. If tsunami earthquakes are unusually shallow and/or involve sediments of low seismic wave speed, this is probably a poor approximation of the actual structure near the source region (Satake

and Tanioka 1999; Geist and Bilek 2001). Some authors have attempted to rectify this error by using moment or slip distributions determined by seismic inversions in a structural model with significantly reduced shear modulus, more appropriate for the shallow trench region, and forward modeling the tsunami waves (e.g., Geist and Bilek 2001). This approach, however, is not satisfactory because a seismic inversion for moment or slip using this more appropriate rigidity model would produce a different distribution of moment or slip. Thus, a simple “correction” for the use of an inappropriate value of rigidity cannot be carried out after the rupture model has already been computed. If available, a correct rigidity model should be part of the seismic inversion process itself. To do this correctly, Green’s functions should be computed for a three-dimensional (or possibly two dimensional, if the velocity structure is relatively uniform in the trench-parallel direction) velocity structure of the shallow subduction zone and incorporated in the modeling of the seismic waveforms. However, such sophisticated models are currently only available for very few subduction zones, and the computational power required for these calculations (for body wave frequencies) is substantial. Unfortunately, whichever approach is chosen to go from recordings of seismic waveforms of tsunami earthquakes to the prediction of tsunami wave heights, a good model of the velocity and elastic properties of the shallow subduction zone is an unavoidable requirement.

Even when the moment distribution or moment magnitude of a tsunami earthquake has been determined using an appropriate model for the velocity and elastic parameters, there exists also the issue of enhanced tsunami excitation in material with weaker elastic properties, such as sedimentary layers. Modeling suggests that an event for which 10 % of the moment is in sediments generates a tsunami ten times larger than its seismic moment would suggest (Okal 1988) mainly because the slip in this material would be much greater than that for the same seismic moment in a stronger material (moment is the product of slip, area, and rigidity after all). Therefore, the moment of a tsunami earthquake, even if

determined correctly, may not directly reflect its tsunamigenic potential when low-velocity sediments are present in the rupture zone. Since tsunami wave heights are mainly determined by the vertical displacement of the ocean floor, which in turn is primarily controlled by the slip on the fault plane, the slip (distribution) is more directly indicative of tsunamigenic potential. Since variations in shear modulus of a factor of five are not uncommon in shallow subduction zones (Geist and Bilek 2001), earthquakes with similar moments can result in substantially different slip models and tsunami excitation.

Shallow Depth of Slip Causes Relatively Great Displacement of Ocean Floor

Shallower earthquakes produce greater and shorter wavelength vertical displacement of the ocean floor and thus greater and shorter wavelength tsunami waves right above the source region. However, higher-frequency waves travel more slowly than longer-period waves, and, after a few hundred or thousand kilometers of travel, they drift to the back of the wave train and do not contribute to the maximum amplitude. Beyond about 2,000 km distance, any earthquake at a depth less than 30 km appears to be equally efficient in tsunamigenesis (Ward 2002 and Fig. 6). Therefore, the exact depth of the slip in a shallow earthquake is not a key factor in determining its teleseismic tsunami wave heights. Although the teleseismic tsunami wave heights for a shallow slip event may not be significantly greater in amplitude than those for a somewhat deeper slip event (Ward 2002), at local and regional distances the depth of the slip is an important factor. Therefore, for tsunami earthquakes, which have anomalously great tsunami height at mainly local and regional distances close to the rupture, the depth of the rupture should be a significant factor in their tsunamigenesis (see Fig. 6). Furthermore, modeling of shallow subduction zone earthquakes using a specific crack model (Geist and Dmowska 1999) indicates that a rupture intersecting the free surface results in approximately twice the average slip. However, under the assumption of other specific frictional

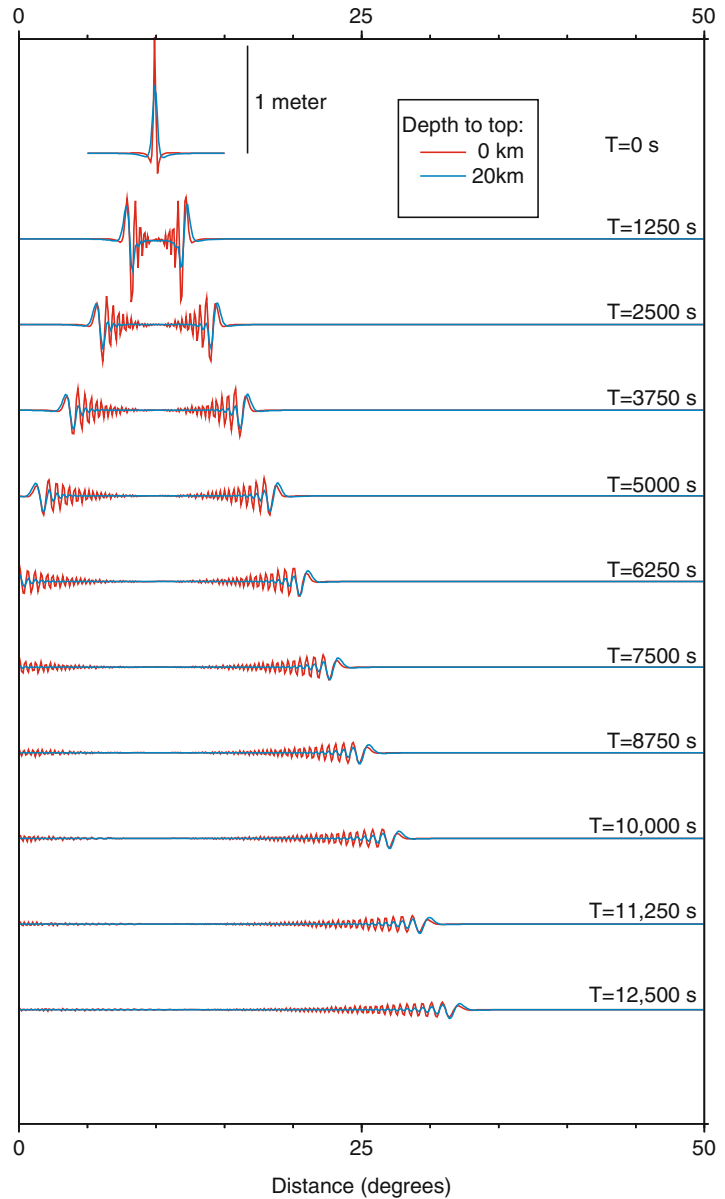
and crack models, the modeling of subduction zone earthquakes by Wang and He (2007) produces slip models that have less vertical displacement of the ocean floor when the slip reaches the surface, due to a different slip distribution and the curvature of the subduction interface.

Shallow Fault Dip May Lead to Underestimation of Slip from Seismic Waves

It is notoriously difficult to resolve moment, M_0 , and dip, δ , independently for shallow thrust earthquakes. The excitation functions of Rayleigh waves for shallow thrust earthquakes show that only the product of dip and moment (more precisely $M_0 \sin(2\delta)$) can be resolved (Kanamori and Given 1981) when using these surface waves in inversions for a source mechanism. If Love wave data are also included in the inversion, it may be possible to add constraints by concentrating on fitting the amplitudes of those Love waves recorded at azimuths corresponding to the along strike direction (Ji 2006a); however, this approach may be complicated by directivity effects. The polarity of body waves can only be used to constrain the focal mechanism of an earthquake at a limited range of incidence angles and thus also cannot provide any additional constraints on the dip of the shallowly dipping plane, unless assumptions are made about the rake angle. It is thus possible to (severely) underestimate the amount of slip in the earthquake, if the dip of the mechanism is poorly constrained and the inversion leads to a value for dip that is too high. This could be particularly important for very shallow subduction earthquakes, since the dip is expected to be small for these events. Thus, a difference between a dip of 3 or 6° in a CMT solution may not seem significant, but it could lead to a difference in moment (and thus slip) of a factor of two. To illustrate the importance of this issue for very shallowly dipping thrust events, we show in Fig. 7 three different homogeneous slip models that will produce similar surface waves because the product of their slip and dip (and thus moment) is identical. However, the vertical deformation of the ocean floor and the tsunami

Tsunami Earthquakes,

Fig. 6 Cross sections of an expanding tsunami from a M7.5 thrust earthquake. The fault strikes north south (into the page) and the sections are taken east west. Elapsed time in seconds is given at the left and right sides. *Red lines* are for a fault that breaks the surface and *blue lines* for a fault with its top at a depth of 20 km. Deeper earthquakes make smaller and longer wavelength tsunamis at relatively short distances



waveforms resulting from these three different earthquakes would be significantly different in amplitude.

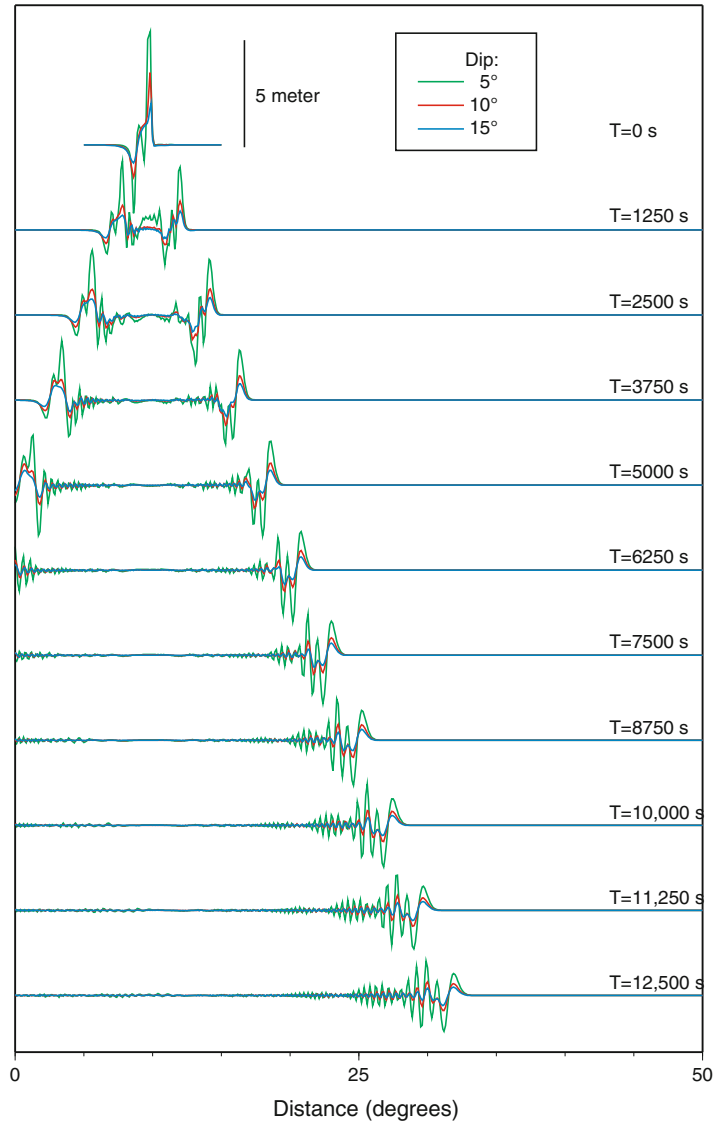
Horizontal Deformation of the Ocean Floor May Lead to Great Displacement of Water Yet Is Neglected in Tsunami Modeling

Most tsunami modelers consider only a water surface displacement identical to the vertical deformation of the ocean bottom due to faulting

when computing the tsunami height resulting from an earthquake and neglect the effect of horizontal deformation. However, when the tsunami source is located close to a steep slope and the horizontal displacement is large relative to the vertical displacement, which is generally the case for tsunami earthquakes due to their mechanism and shallow depth, the effect of horizontal deformation may become significant (Tanioka and Satake 1996b).

Tsunami Earthquakes,

Fig. 7 Tsunami wave heights as a function of time for three different slip models for a shallow thrust earthquake, which will produce similar surface wave recordings (because the product of dip and moment is held constant). Thus, it is difficult to resolve between these different models using an inversion of surface waves, yet they produce very different vertical displacement of the ocean floor and thus very different tsunamis



Furthermore, it has been suggested that the lateral collision force of a continental slope into the ocean due to faulting could also play an important part in the tsunamigenesis of these events (Song et al. 2006). This type of dynamic excitation of tsunami waves would be particularly important for very shallowly dipping, shallow, thrust events, which would have a large component of horizontal motion.

Subduction of Bathymetric Features May Enable Seismic Slip in Usually Aseismic Region and Be Related to Unusual Aftershock Sequences

Sandbox experiments show the pervasive influence on the geomorphology of the shallow subduction zone margin when a seamount on the subducting plate is being subducted (Dominguez et al. 2000). Subduction and underplating of relatively undeformed and water-laden sediments

beneath the rear part of the margin could, together with the dense fracture network generated by seamount subduction, modify the fluid pressure and introduce significant variations of the effective basal friction and thus the local mechanical plate coupling. More directly, subduction of a seamount may increase the normal stress across the subduction interface and hence enhance seismic coupling (Scholz and Small 1997; Cloos 1992). Unusual earthquakes have been documented in regions where ridges, seamounts, or other bathymetric features are being subducted (e.g. Chung and Kanamori 1978; Kodaira et al. 2000), and investigations of rupture characteristics of large underthrusting earthquakes provide evidence that seamounts can be subducted to seismogenic depths and that variations in seafloor bathymetry of the subducting plate may strongly influence the earthquake rupture process (Bilek et al. 2003; Robinson et al. 2006). In the case of the Java tsunami earthquake of 1994, the bathymetry of the area landwards of the trench suggests that a local high is in the process of being subducted close to the area of maximum slip (Abercrombie et al. 2001, Fig. 5). A bathymetric high in the form of the Roo Rise can also be found just seaward of the trench region. In the bathymetry of the area around the 2006 Java event, no such pronounced local feature can be found (Fig. 5), but the regional bathymetry south of the Java trench region is distinguished by an overall rough character. Similarly, the Nicaragua subduction zone is characterized by a highly developed horst-and-graben structure in the subducting plate, but no large-scale features, like a subducting seamount, are obvious. However, in case of the Peru earthquakes, both the 1996 and 1960 tsunami earthquakes occur at the intersection of the trench with major topographic features on the Nazca plate: the Mendaña fracture zone and the Trujillo trough, respectively (Okal and Newman 2001). Thus, subduction of either pronounced local bathymetric features or more regional seafloor roughness or horst-and-graben structures, which may modify the local coupling between subducting and overriding plates, has been

documented in or near the rupture zone of many tsunami earthquakes.

Accretionary Prism: Uplift, Slides, or Splay Faulting May Displace a Relatively Great Volume of Water

Tsunami earthquakes may involve seismic slip along the normally aseismic basal decollement of the accretionary prism (Pelayo and Wiens 1992; Tanioka and Satake 1996a). Sediments near the toe of an inner trench slope may be scraped off by a large horizontal movement over the decollement due to an earthquake and thus cause an additional inelastic uplift, which could have a large effect on tsunami generation (for the 1896 Sanriku earthquake, Tanioka and Seno 2001a; for the 1946 Aleutian Islands earthquake, Tanioka and Seno 2001b).

The existence of splay faulting, which would be more effective in exciting tsunamis due to their steeper dip, within the accretionary prism itself has been suggested to be a cause of the large tsunami excitation for the 1994 Java earthquake (Abercrombie et al. 2001), the 2004 Sumatra megathrust event (Lay et al. 2005), and the 1963/1975 Kurile earthquakes (Fukao 1979). Splay faulting can further promote extensive vertical deformation of the ocean floor, and hence large tsunamis, through partitioning or branching of a rupture upwards from the interface along multiple splay faults leading up to the surface (Fukao 1979; Park et al. 2002).

Presence of Fluids Influences Seismic Behavior

In subduction zones, fluids expelled from the subducting plate play an important role in many different subduction-related phenomena such as volcanism, metamorphism, and seismogenesis. Zones of high pore fluid pressure in the shallow subduction zone would change the effective normal stress significantly, possibly extending the region in which seismic slip is possible to shallower depths and generate slip of a slow nature. The presence of such zones has been

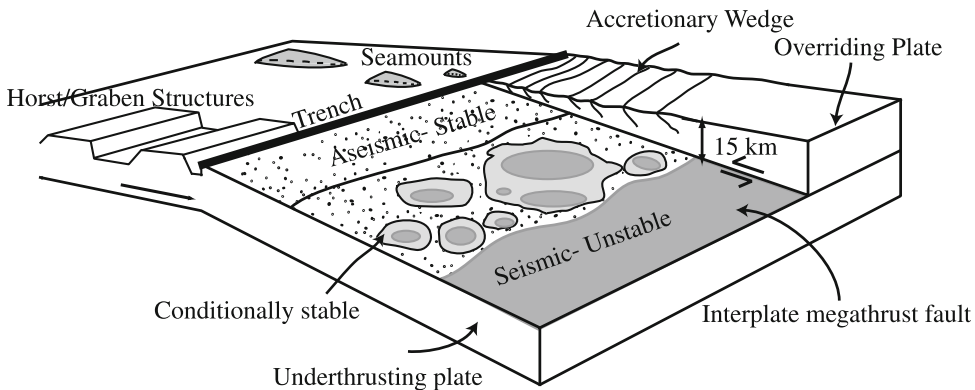
suggested to be related to the occurrence of silent slip in the Nankai trough region (Kodaira et al. 2004). In the case of tsunami earthquakes, the presence of large zones of elevated fluid pressure has been proposed to cause fairly rapid seismic slip close to the trench axis, following the breakage of asperities (Seno 2002).

A Model for Tsunami Earthquakes

Most explanations for the slowness of tsunami earthquakes involve the presence of low-velocity, low-strength, and low-rigidity sediments in the accretionary prism and between the overriding and subducting plate in the shallow subduction zone. Rupture through these slow sediments is thought to promote a slow rupture velocity. In Scholz's (1990, 1998) model of the typical subduction zone (see Fig. 8 for an interpretation of this model from Bilek and Lay 2002), three possible stability regimes exist. In the stable zone, seismic slip cannot be supported and aseismic creep releases all strain. In the unstable zone episodic seismic slip occurs. In the conditionally stable zone, in between these two zones, slip can be abrupt if it experiences loading from a nearby slip patch. This conditionally stable zone may be very heterogeneous due to roughness of the thrust fault (Tanioka et al. 1997) creating isolated asperities or due to permeability changes

(Pacheco et al. 1993) from the subduction of low-permeability materials or the presence of fluids (Seno 2002). Tsunami earthquakes may represent a rupture of one or several large “unstable” asperities, which then propagates for a significant distance in the conditionally stable sedimentary materials (Polet and Kanamori 2000; Bilek and Lay 2002; Hu and Wang 2008; Lay et al. 2011, 2012).

Alternatively, but consistently, we could interpret tsunami earthquakes in the context of fracture mechanics as displaying a lack of radiated energy and a low rupture speed in a high- G_C (fracture energy) environment (Ammon et al. 2006; Venkataraman and Kanamori 2004). As stated above, if these ruptures also involve localized patches of relatively strong unstable friction (that would be associated with high rupture speed and low G_C), it would allow the rupture to propagate seismically, instead of as a continuous creep process. From this point of view, tsunami earthquakes dissipate a large amount of energy during the fracture process and are left with little energy to radiate. It is possible that the highly faulted trench and deformed sediments result in larger energy dissipation during failure due to an excessive amount of branching and bifurcation of cracks which gives rise to inelastic behavior and hence a large dissipation of energy (Barragan et al. 2001; Venkataraman and Kanamori 2004), possibly involving the branching of the rupture



Tsunami Earthquakes, Fig. 8 Cartoon illustrating frictional conditions of the subduction interface between subducting and overriding plate. Individual unstable sliding contact areas (dark gray) can provide the nucleation

sites for rupture in the shallow subduction zone environment, which is typically a stable (stippled) or conditionally stable (light gray) frictional region (From Bilek and Lay (2002))

into multiple splay faults in the accretionary prism.

Tsunami earthquakes therefore would represent slip at unusually shallow depths that would typically be dominated by creep processes. Their nucleation would be made possible through the existence of localized asperities or patches of unstable friction in a typically stable or conditionally stable region. The presence of compartments of elevated fluid pressure may aid in the propagation of the seismic slip by creating zones of nearly zero friction surrounding the asperities (Seno 2002). These asperities may be created by the subduction of bathymetric features like seamounts or ridges or by the broken-up nature of the subducting plate itself, creating a horst-and-graben system, which would act as buckets for sediment subduction (Tanioka et al. 1997; Polet and Kanamori 2000). The stress release on these asperities would be near complete, and any additional unloading of stress on the plate interface due to the rupture may occur mostly through creep. This would result in a relatively low number of aftershocks occurring on the interface between overriding and subducting plate. However, the static stress change in the outer rise area would be significant due to the shallow nature of most of the slip, and thus normal faulting outer rise earthquakes would be more likely to be triggered (Dmowska et al. 1996; Taylor et al. 1996; Hilde 1983). The subduction of a bathymetric feature would also likely result in fracturization of the margin (Dominguez et al. 2000) in the overriding plate, thus further promoting the occurrence of intraplate aftershocks in this area of the shallow subduction zone. If the subducting plate itself is highly broken-up and thus contains preexisting weak zones, this may facilitate further faulting within the subducting plate, in particular close to the lower edge of the rupture where the stress change due to the main shock is relatively large.

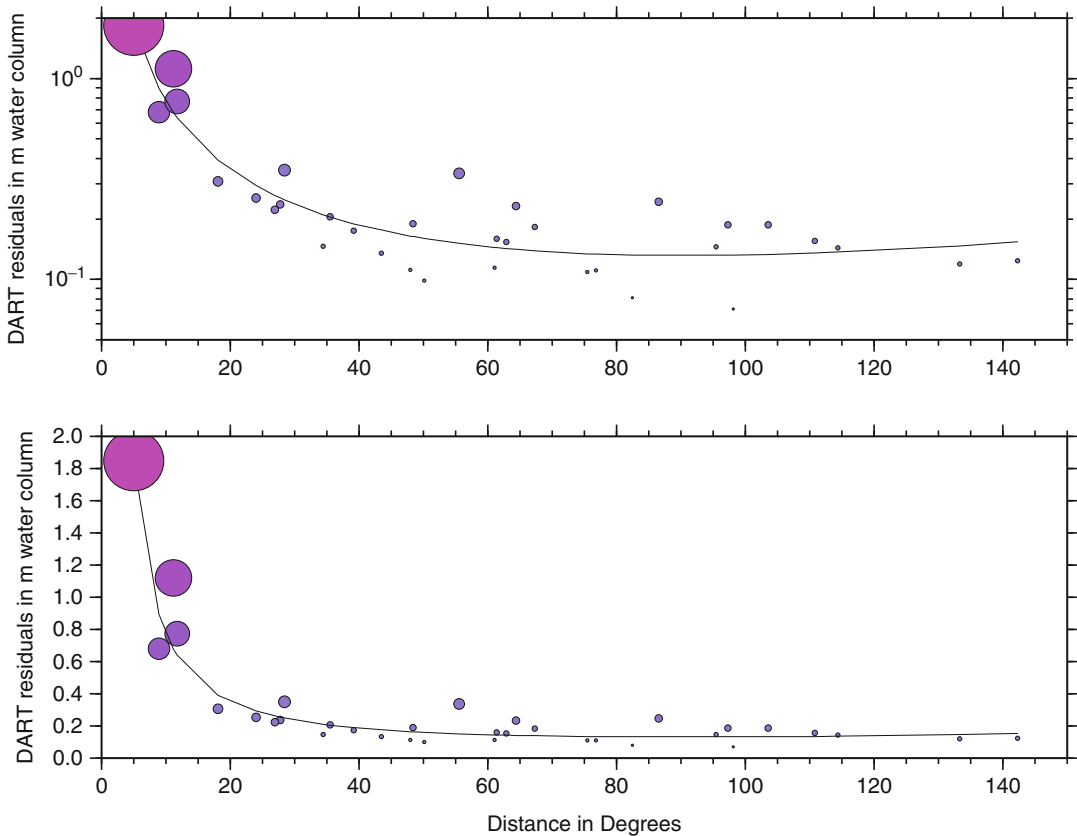
In this model for tsunami earthquakes discussed above, the unusually high effectiveness in the excitation of tsunami waves can be attributed to several factors, with the shallowness of the slip as the main underlying cause. Other important concerns coming into play are the

possible failure of seismological techniques to provide an accurate estimate of the slip due to complexities associated with a very shallowly propagating rupture in a subduction zone, the possible failure of tsunami modeling to determine accurate wave heights due to similar complexities, and the possible involvement of splay faulting or uplift of sediments near the trench in the accretionary prism.

Future Directions

The Sumatra-Andaman and Tohoku earthquakes and tsunami renewed interest in the development of near real-time methods to estimate the true size of large earthquakes and the tsunamis that might follow them and the installation of instrumentation that will facilitate these measurements. Because the time between the earthquake and the arrival of the first tsunami waves at the local coastline is short, it is still unclear how effective these types of early warning systems are for saving lives at short distances from the tsunami source, but they will be useful at large distances.

New technologies and surveys will enhance our knowledge of the geomorphology and velocity structure of the shallow subduction zone. Ocean bottom seismometers, tide gauges, buoys, and other seafloor monitoring devices will provide high-quality data, which will enable us to place better constraints on where exactly the slip in shallow earthquakes occurs and in what tectonic environment. The Deep-ocean Assessment and Reporting of Tsunamis (DART) system provides open-ocean tsunami wave-height data that will allow for the calculation of reliable tsunami magnitudes for great earthquakes (Fig. 9), once the database of events is sufficient to carry out a calibration. Although tsunami earthquakes occur relatively infrequently and thus may be difficult to capture, comprehensive characterizations of their rupture processes placed in the context of detailed three-dimensional models of the shallow subduction zones they occurred in will be an important next step in understanding their unusual seismo- and tsunamigenic processes.



Tsunami Earthquakes, Fig. 9 DART measurements of open-ocean maximum wave height for 2011 Tohoku earthquake, corrected for tides, as provided by the

National Geophysical Data Center (<http://www.ngdc.noaa.gov/hazard/DARTData.shtml>). Line indicates exponential fit to data

Bibliography

Primary Literature

- Abe K (1979) Size of great earthquakes of 1873–1974 inferred from tsunami data. *J Geophys Res* 84:1561–1568
- Abe K (1989) Quantification of tsunamigenic earthquakes by the M_i scale. *Tectonophysics* 166:21–34
- Abercrombie RE, Antolik M, Felzer K, Ekstrom G (2001) The 1994 Java tsunami earthquake—slip over a subducting seamount. *J Geophys Res* 106:6595–6608
- Ammon CJ, Ji C, Thio HK, Robinson D, Ni S, Hjorleifsdottir V, Kanamori H, Lay T, Das S, Helmberg D, Ichinose G, Polet J, Wald D (2005) Rupture process of the 2004 Sumatra-Andaman earthquake. *Science* 308:1133. <https://doi.org/10.1126/science.1112260>
- Ammon CJ, Kanamori H, Lay T, Velasco AA (2006) The 17 July 2006 Java tsunami earthquake. *Geophys Res Lett* 33:24. <https://doi.org/10.1029/2006GL028005>

- Barragan BE, Giaccio GM, Zerbino RL (2001) Fracture and failure of thermally damaged concrete under tensile loading. *Mater Struct* 34:312–319
- Beck SL, Ruff LJ (1987) Rupture process of the great 1963 Kuril Islands earthquake sequence: asperity interaction and multiple event rupture. *J Geophys Res* 92:14123–14138
- Bilek SL, Engdahl ER (2007) Rupture characterization and aftershock relocations for the 1994 and 2006 tsunami earthquakes in the Java subduction zone. *Geophys Res Lett* (in press)
- Bilek SL, Lay T (2002) Tsunami earthquakes possibly widespread manifestations of frictional conditional stability. *Geophys Res Lett* 29:18–1. <https://doi.org/10.1029/2002GL01521>
- Bilek SL, Schwartz SY, Deshon HR (2003) Control of seafloor roughness on earthquake rupture behavior. *Geology* 31:455–458. [https://doi.org/10.1130/0091-7613\(2003\)031](https://doi.org/10.1130/0091-7613(2003)031)

- Bourgeois J, Petroff C, Yeh H, Titov V, Synolakis CE, Benson B, Kuroiwa J, Lander J, Norabuena E (1999) Geologic setting, field survey and modeling of the Chimbote, Northern Peru, tsunami of 21 February 1996. *Pure Appl Geophys* 154:513–540
- Brown DL (1964) Tsunami activity accompanying the Alaskan earthquake of 27 March 1964. U.S. Army Engr. Dist, Alaska, 20 pp
- Brune JN (1970) Tectonic stress and spectra of seismic shear waves from earthquakes. *J Geophys Res* 75:4997–5009
- Chen T, Newman AV, Feng L, Fritz HM (2009) Slip distribution from the 1 April 2007 Solomon Islands earthquake: a unique image of near-trench rupture. *Geophys Res Lett* 36:L16307, <https://doi.org/10.1029/2009GL039496>
- Chung WY, Kanamori H (1978) Subduction process of a fracture zone and aseismic ridges – the focal mechanism and source characteristics of the New Hebrides earthquake of 1969 January 19 and some related events. *Geophys J Int* 54(1):221–240. <https://doi.org/10.1111/j.1365-246X.1978.tb06764.x>
- Cloos M (1992) Thrust-type subduction-zone earthquakes and seamount asperities; a physical model for seismic rupture. *Geology* 20:601–604
- Dengler L, Uslu B, Barberopoulou A, Yim SC, Kelly A (2009) The November 15, 2006 Kuril Islands-generated tsunami in Crescent City, California. *Pure Appl Geophys* 166:37–53. <https://doi.org/10.1007/s00024-008-0429-2>
- Dmowska R, Zheng G, Rice JR (1996) Seismicity and deformation at convergent margins due to heterogeneous coupling. *J Geophys Res* 101:3015–3029
- Dominguez S, Malavieille J, Lallemand SE (2000) Deformation of accretionary wedges in response to seamount subduction: insight from sandbox experiments. *Tectonics* 19:182–196
- Engdahl ER, Villaseñor A (2002) Global seismicity: 1900–1999. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) *International handbook of earthquake and engineering seismology*, part a, chapter 41. Academic, San Diego, pp 665–690
- Fritz HM, Kongko W, Moore A, McAdoo B, Goff JR, Harbitz C, Uslu B, Kalligeris N, Suteja D, Kalsum K, Titov V, Gusman A, Latief H, Santoso E, Sujoko S, Djulkarnaen D, Sunendar H, Synolakis CE (2007) Extreme runup from the 17 July 2006 Java tsunami. *Geophys Res Lett* 34:L12602. <https://doi.org/10.1029/2007GL029404>
- Fryer GJ, Watts P, Pratson LF (2004) Source of the great tsunami of 1 April 1946: a landslide in the upper Aleutian forearc. *Geology* 203:201–218
- Fujii Y, Satake K (2006) Source of the July 2006 West Java tsunami estimated from tide gauge records. *Geophys Res Lett* 33. <https://doi.org/10.1029/2006GL028049>
- Fukao Y (1979) Tsunami earthquakes and subduction processes near deep-sea trenches. *J Geophys Res* 84:2303–2314
- Geist E (2000) Origin of the 17 July 1998 Papua New Guinea tsunami: earthquake or landslide? *Seismol Res Lett* 71:344–351
- Geist E, Bilek SL (2001) Effect of depth-dependent shear modulus on tsunami generation along subduction zones. *Geophys Res Lett* 28:1315–1318
- Geist EL, Dmowska R (1999) Local tsunamis and distributed slip at the source. *Pure Appl Geophys* 154:485–512
- Hara T (2006) Determination of earthquake magnitudes using duration of high-frequency energy radiation and maximum displacement amplitudes: application to the July 17, 2006 Java earthquake and other tsunami earthquakes. *EOS Trans Am Geophys Union* 87(52), Fall Meet. Suppl., Abstract S21A–0132
- Hatori T (1967) The generating area of Sanriku tsunami of 1896 and its comparison with the tsunami of 1933. *J Seismol Soc Jpn* 20:164–170
- Heinrich P, Schindele F, Guibourg S, Ihmlé PF (1998) Modeling of the February 1996 Peruvian tsunami. *Geophys Res Lett* 25:2687–2690
- Hidayat D, Barker JS, Satake K (1995) Modeling the seismic source and tsunami generation of the December 12, 1992 Flores island, Indonesia, earthquake. *Pure Appl Geophys* 144:537–554
- Hilde TWC (1983) Sediment subduction versus accretion around the Pacific. *Tectonophysics* 99:381–397
- Hill EM, Borrero JC, Huang Z, Qiu Q, Banerjee P, Natawidjaja DH, Elosegui P, Fritz HM, Suwargadi BW, Pranantyo IR, Li L, Macpherson KA, Skanavis V, Synolakis CE, Sieh K (2012) The 2010 M_w 7.8 Mentawai earthquake: very shallow source of a rare tsunami earthquake determined from tsunami field survey and near-field GPS data. *J Geophys Res Solid Earth* 117:B06402
- Hu Y, Wang KL (2008) Coseismic strengthening of the shallow portion of the subduction fault and its effects on wedge taper. *J Geophys Res Solid Earth* 113: B12411. <https://doi.org/10.1029/2008jb005724>
- Ide S, Imamura F, Yoshida Y, Abe K (1993) Source characteristics of the Nicaraguan tsunami earthquake of September 2, 1992. *Geophys Res Lett* 20:863–866
- Ihmlé PF, Gomez JM, Heinrich P, Guibourg S (1998) The 1996 Peru tsunamigenic earthquake: broadband source process. *Geophys Res Lett* 25:2691–2694
- Imamura F, Gica E, Takahashi T, Shuto N (1995) Numerical simulation of the 1992 Flores tsunami: interpretation of tsunami phenomena in northeastern Flores Island and damage at Babi Island. *Pure Appl Geophys* 144:555–568
- Ji C (2006a) Resolving the trade-off between the seismic moment and fault dip of large subduction earthquakes and its impact on tsunami excitation. *Tsunami Sources Workshop*
- Ji C (2006b) A comparison study of 2006 Java earthquake and other tsunami earthquakes. *EOS Trans Am Geophys Union* 87(52), Fall Meet. Suppl., Abstract

- Ji C, Zeng Y, Song AT (2007) Rupture process of the 2006 M_w 8.3 Kuril Island Earthquake inferred from joint inversion of teleseismic body and surface waves, SSA meeting
- Johnson JM, Satake K (1997) Estimation of seismic moment and slip distribution of the April 1, 1946, Aleutian tsunami earthquake. *J Geophys Res* 102:11765–11774
- Kanamori H (1972) Mechanism of tsunami earthquakes. *Phys Earth Planet Inter* 6:346–359
- Kanamori H (1993) W phase. *Geophys Res Lett* 20:1691–1694
- Kanamori H, Given JW (1981) Use of long-period surface waves for rapid determination of earthquake-source parameters. *Phys Earth Planet Inter* 27:8–31
- Kanamori H, Kikuchi M (1993) The 1992 Nicaragua earthquake – a slow tsunami earthquake associated with subducted sediments. *Nature* 361:714–716
- Kikuchi M, Kanamori H (1995) Source characteristics of the 1992 Nicaragua tsunami earthquake inferred from teleseismic body waves. *Pure Appl Geophys* 144:441–453
- Kodaira S, Takahashi N, Nakanishi A, Miura S, Kaneda Y (2000) Subducted seamount imaged in the rupture zone of the 1946 Nankaido earthquake. *Science* 289: 104–106. <https://doi.org/10.1126/science.289.5476.104>
- Kodaira S, Iidaka T, Kato A, Park JO, Iwasaki T, Kaneda Y (2004) High pore fluid pressure may cause silent slip in the Nankai Trough. *Science* 304:1295–1298. <https://doi.org/10.1126/science.1096535>
- Koper KD, Hutko AR, Lay T, Ammon CJ, Kanamori H (2011) Frequency-dependent rupture process of the 2011 M_w 9.0 Tohoku earthquake: comparison of short-period P wave back projection images and broadband seismic rupture models. *Earth Planets Space* 63:599–602
- Kulm LD, Prince RA, French W, Johnson S, Masias A (1981) Crustal structure and tectonics of the central Peru continental margin and trench. In: Kulm LD, Dymond J, Dasch EJ, Hussong DM (eds) *Nazca plate: crustal formation and Andean Convergence*. *Geol Soc Am Mem* 154:445–468
- Lavigne F, Gomes C, Giffo M, Wassmer P, Hoebreck C, Mardiatno D, Priyono J, Paris R (2007) Field observations of the 17 July 2006 Tsunami in Java. *Nat Hazards Earth Syst Sci* 7:177–183
- Lay T, Kanamori H, Ammon CJ, Nettles M, Ward SN, Aster RA, Beck SL, Bilek BL, Brudzinski MR, Butler R, DeShon HR, Ekström G, Satake K, Sipkin S (2005) The great Sumatra-Andaman earthquake of 26 December 2004. *Science* 308. <https://doi.org/10.1126/science.1112250>
- Lay T, Ammon CJ, Kanamori H, Yamazaki Y, Cheung KF, Hutko AR (2011) The 25 October 2010 Mentawai tsunami earthquake (M_w 7.8) and the tsunami hazard presented by shallow megathrust ruptures. *Geophys Res Lett* 38:L06302. <https://doi.org/10.1029/2010GL046552>
- Lay T, Kanamori H, Ammon CJ, Koper KD, Hutko AR, Ye L, Yue H, Rushing TM (2012) Depth-varying rupture properties of subduction zone megathrust faults. *J Geophys Res* 117:B04311. <https://doi.org/10.1029/2011JB009133>
- Lopez AM, Okal EA (2006) A seismological reassessment of the source of the 1946 Aleutian ‘tsunami’ earthquake. *Geophys J Int* 165(3):835–849. <https://doi.org/10.1111/j.1365-246X.2006.02899.x>
- Masson DG, Parson LM, Milsom J, Nichols G, Sikumbang N, Dwiyoanto B, Kallagher H (1990) Subduction of seamounts at the Java trench – a view with long-range sidescan sonar. *Tectonophysics* 185:51–65
- McAdoo BG, Capone MK, Minder J (2004) Seafloor geomorphology of convergent margins: implications for Cascadia seismic hazard. *Tectonics* 23. <https://doi.org/10.1029/2003TC001570>
- Mori J, Mooney WD, Afnimar Kurniawan S, Anaya AI, Widiyantoro S (2007) The 17 July 2006 tsunami earthquake in West Java, Indonesia. *Seismol Res Lett* 78:291
- Newman AV, Okal EA (1998) Teleseismic estimates of radiated seismic energy: the E/M_0 discriminant for tsunami earthquakes. *J Geophys Res* 103: 26885–26898
- Newman AV, Hayes G, Wei Y, Convers J (2011) The 25 October 2010 Mentawai tsunami earthquake, from real-time discriminants, finite-fault rupture, and tsunami excitation. *Geophys Res Lett* 38(5). <https://doi.org/10.1029/2010GL046498>
- Okal EA (1988) Seismic parameters controlling far-field tsunami amplitudes: a review. *Nat Hazards* 1:67–96
- Okal EA, Newman AV (2001) Tsunami earthquakes: the quest for a regional signal. *Phys Earth Planet Inter* 124:45–70
- Okamoto T, Takenaka H (2009) Waveform inversion for slip distribution of the 2006 Java tsunami earthquake by using 2.5D finite-difference Green’s function. *Earth Planets Space* 61:e17–e20
- Pacheco JF, Sykes LR, Scholz CH (1993) Nature of seismic coupling along simple plate boundaries of the subduction type. *J Geophys Res* 98:14,133–14,159
- Park JO, Tsuru T, Kodaira S, Cummins PR, Kaneda Y (2002) Splay fault branching along the Nankai subduction zone. *Science* 297:1157–1160
- Pelayo AM, Wiens DA (1990) The November 20, 1960 Peru tsunami earthquake: source mechanism of a slow event. *Geophys Res Lett* 17:661–664
- Pelayo AM, Wiens DA (1992) Tsunami earthquakes – slow thrust-faulting events in the accretionary wedge. *J Geophys Res* 97:15,321–15,337
- Polet J, Kanamori H (2000) Shallow subduction zone earthquakes and their tsunamigenic potential. *Geophys J Int* 142:684–702. <https://doi.org/10.1046/j.1365-246x.2000.00205.x>
- Polet J, Thio HK (2003) The 1994 Java tsunami earthquake and its “Normal” aftershocks. *Geophys Res Lett* 30:27–1. <https://doi.org/10.1029/2002GL016806>
- Robinson DP, Das S, Watts AB (2006) Earthquake rupture stalled by a subducting fracture zone. *Science* 312:1203–1205. <https://doi.org/10.1126/science.1125771>
- Satake K (1994) Mechanics of the 1992 Nicaragua tsunami earthquake. *Geophys Res Lett* 21:2519–2522
- Satake K, Kanamori H (1991) Abnormal tsunamis caused by the June 13, 1984, Torishima, Japan, earthquake. *J Geophys Res* 96:19933–19939

- Satake K, Tanioka Y (1999) Sources of tsunami and tsunamigenic earthquakes in subduction zones. *Pure Appl Geophys* 154:467–483. <https://doi.org/10.1007/s000240050240>
- Satake K, Bourgeois J, Abe K, Abe K, Tsuji Y, Imamura F, Iio Y, Katao H, Noguera E, Estrada F (1993) Tsunami field survey of the 1992 Nicaragua earthquake. *EOS Trans Am Geophys Union* 74:156–157
- Satake K, Fujii Y, Harada T, Namegaya Y (2013) Time and space distribution of coseismic slip of the 2011 Tohoku earthquake as inferred from tsunami waveform data. *Bull Seismol Soc Am* 103:1473–1492
- Scholz C (1990) *The mechanics of earthquakes and faulting*. Cambridge University Press, New York
- Scholz C (1998) Earthquakes and friction laws. *Nature* 391:37–42
- Scholz CH, Small C (1997) The effect of seamount subduction on seismic coupling. *Geology* 25:487–490
- Seno T (2002) Tsunami earthquakes as transient phenomena. *Geophys Res Lett* 29:10. <https://doi.org/10.1029/2002GL014868>
- Seno T, Hirata K (2007) Did the 2004 Sumatra–Andaman earthquake involve a component of tsunami earthquakes? *Bull Seismol Soc Am* 97:S296–S306. <https://doi.org/10.1785/0120050615>
- Shapiro NM, Singh SK, Pacheco J (1998) A fast and simple diagnostic method for identifying tsunamigenic earthquakes. *Geophys Res Lett* 25:3911–3914
- Shimazaki K, Geller RJ (1977) Source process of the Kurile Islands tsunami earthquake of June 10, 1975. *EOS Trans Am Geophys Union* 58:446
- Simons M, Minson SE, Sladen A, Ortega F, Jiang J, Owen SE, Meng L, Ampuero J-P, Wei S, Chu R, Helmberger DV, Kanamori H, Hetland E, Moore AW, Webb FH (2011) The 2011 magnitude 9.0 Tohoku–Oki earthquake: mosaicking the megathrust from seconds to centuries. *Science* 332:1421–1425
- Song Y, Fu L, Zlotnicki V, Ji C, Hjorleifsdottir V, Shum C, Yi Y (2006) Horizontal motions of faulting dictate the 26 December 2004 tsunami genesis. *EOS Trans Am Geophys Union* 87(52), Fall Meet. Suppl., Abstract U53C-02
- Synolakis C, Imamura F, Tsuji Y, Matsutomi H, Tinti S, Cook B, Chandra YP, Usman M (1995) Damage, conditions of East Java tsunamis of 1994 analyzed. *Eos* 76(26):257
- Synolakis CE, Bardet JP, Borrero JC, Davies HL, Okal EA, Silver EA, Sweet S, Tappin DR (2002) The slump origin of the 1998 Papua New Guinea tsunami. *Proc R Soc A Math Phys Eng Sci* 458(2020):763–789. <https://doi.org/10.1098/rspa.2001.0915>
- Tanioka Y, Satake K (1996a) Fault parameters of the 1896 Sanriku tsunami earthquake estimated from tsunami numerical modeling. *Geophys Res Lett* 23:1549–1552
- Tanioka Y, Satake K (1996b) Tsunami generation by horizontal displacement of ocean bottom. *Geophys Res Lett* 23:861–864
- Tanioka Y, Seno T (2001a) Sediment effect on tsunami generation of the 1896 Sanriku tsunami earthquake. *Geophys Res Lett* 28:3389–3392
- Tanioka Y, Seno T (2001b) Detailed analysis of tsunami waveforms generated by the 1946 Aleutian tsunami earthquake. *Nat Hazards Earth Syst Sci* 1:171–175
- Tanioka Y, Ruff L, Satake K (1997) What controls the lateral variation of large earthquake occurrence along the Japan trench. *Island Arc* 6:261–266
- Taylor MAJ, Zheng G, Rice JR, Stuart WD, Dmowska R (1996) Cyclic stressing and seismicity at strong coupled subduction zones. *J Geophys Res* 101:8363–8381
- Tsuboi S (2000) Application of M_{wp} to tsunami earthquake. *Geophys Res Lett* 27:3105–3108
- Tsuji Y, Imamura F, Matsutomi H, Synolakis CE, Nanang PT, Jumadi, Harada S, Han SS, Arai K, Cook B (1995) Field survey of the East Java earthquake and Tsunami of June 3, 1994. *Pure Appl Geophys* 144(3/4):839–854
- Venkataraman A, Kanamori H (2004) Observational constraints on the fracture energy of subduction zone earthquakes. *J Geophys Res* 109. <https://doi.org/10.1029/2003JB002549>
- Wang K, He J (2007) Effects of frictional behaviour and geometry of subduction fault on coseismic seafloor deformation. Submitted to *Bull Seismol Soc Am*
- Ward SN (2002) Tsunamis. In: Meyers RA (ed) *The encyclopedia of physical science and technology*, vol 17. Academic, pp 175–191
- Wiens D (1989) Bathymetric effects on body waveforms from shallow subduction zone earthquakes and application to seismic processes in the Kurile Trench. *J Geophys Res* 94:2955–2972
- Yue H, Lay T, Rivera L, Bai Y, Yamazaki Y, Cheung KF, Hill EM, Sieh K, Kongko W, Muhari A (2014) Rupture process of the 2010 M_w 7.8 Mentawai tsunami earthquake from joint inversion of near-field hr-GPS and teleseismic body wave recordings constrained by tsunami observations. *J Geophys Res*. <https://doi.org/10.1002/2014JB011082>

Books and Reviews

- Bebout G, Kirby S, Scholl D, Platt J (1996) *Subduction from top to bottom*. American geophysical union monograph no. 96. American Geophysical Union, Washington, DC
- Satake K, Imamura F (1995) Tsunamis 1992–1994. *Spec Issue Pure Appl Geophys* 144(3–4):373–890

Tsunamis: Stochastic Models of Occurrence and Generation Mechanisms

Eric L. Geist¹, David D. Oglesby² and Kenny J. Ryan³

¹U.S. Geological Survey, Menlo Park, CA, USA

²Department of Earth Sciences, University of California, Riverside, CA, USA

³Air Force Research Laboratory, Albuquerque, NM, USA

Article Outline

Glossary

Definition of Subject

Introduction

Stochastic Models of Tsunami Generation

Stochastic Models of Tsunami Time Series

Stochastic Models of Tsunami Occurrence

Probabilistic Analysis of Tsunami Hazards

Future Directions

Bibliography

Keywords

Tsunami models · Earthquake models, stochastic process · Tsunami generation · Dynamic earthquake rupture · Tsunami time series · Tsunami occurrence · Tsunami probability

Glossary

Aleatory Uncertainty The uncertainty in seismic and tsunami hazard analysis due to inherent random variability of the quantity being measured. Aleatory uncertainties cannot be reduced by refining modeling or analytical techniques. Modified from Bormann et al. (2013).

Dynamic Earthquake Model A model of time-dependent and spontaneous fault rupture that produces time-dependent 3-D seismic wave and displacement fields on and around the fault, including deformation of the seafloor. A friction evolution equation is specified on the fault during the rupture process.

Dynamic Tsunami Generation Model A model of time-dependent displacement of the water column above the source region computed from the *dynamic earthquake model* (q.v.). It includes propagation of seismic waves in the solid earth and the water column. In contrast to *static and kinematic models* (q.v.), slip is not prescribed; it is computed from time-dependent stress conditions on the fault.

Moment Magnitude (Earthquake) A non-saturating magnitude computed using the scalar *seismic moment* (M_0). It was introduced by Kanamori (1977) via the Gutenberg-Richter magnitude-energy relation. Modified from Bormann et al. (2013).

Initial Tsunami Wavefield The spatially varying field of *tsunami amplitudes* (q.v.) at the start of tsunami propagation, usually in reference to *static tsunami generation models* (q.v.).

Kinematic Earthquake Model A model of time-dependent and prescribed fault slip that produces time-dependent 3-D seismic wave and displacement fields including deformation of the seafloor. In contrast to *dynamic earthquake models* (q.v.), fault slip is not a calculated result of the model.

Kinematic Tsunami Generation Model A model of time-dependent displacement of the water column above the source region that is prescribed from the computed results of either a *kinematic earthquake model* (q.v.) or a *dynamic earthquake model* (q.v.).

Point Process as Applied to Tsunami Events Tsunami events as a random occurrence at isolated points in time and space. Temporal description defined by tsunami

origin or *tsunami arrival time* (q.v.). Spatial description defined by origin location, most often earthquake epicenter, or station location. Tsunami point processes can be marked by tsunami size at an observing station (i.e., a marked point process).

Seismic Coupling (Earthquake) The qualitative ability of a subduction thrust fault to lock and accumulate stress. Strong interplate coupling implies that the fault is locked and capable of accumulating stress, whereas weak coupling implies that the fault is unlocked or only capable of accumulating low stress. A fault with weak interplate coupling could be aseismic or could slip by creep. Also called “interplate coupling.” Modified from Bormann et al. (2013).

Seismic Moment (Earthquake) The magnitude of the component couple of the double couple that is the point force system equivalent to a fault slip in an isotropic elastic body. It is equal to rigidity times the fault slip integrated over the fault plane. It can be estimated from the far-field seismic spectrum or waveform fitting at wavelengths much longer than the source size. It can also be estimated from the near-field seismic, geologic, and geodetic data, measuring consistency among various observations. Also called “scalar seismic moment” to distinguish it from moment tensor. Modified from Bormann et al. (2013).

Source Potency A quantity that characterizes the inelastic property in a tsunami source region. Modified from Bormann et al. (2013).

Subduction Zone A zone of convergence of two lithospheric plates characterized by thrusting of one plate into the Earth’s mantle beneath the other. Processes within the subduction zone bring about melt generation in the mantle wedge and cause buildup of the overlying volcanic arc. Subduction zones (where most of the world’s greatest earthquakes have occurred) are recognized from the systematic distribution of hypocenters of deep earthquakes called Wadati-Benioff zones. Modified from Bormann et al. (2013).

Static Earthquake Model A model of time-independent and prescribed fault slip that

produces a time-independent 3-D deformation field including deformation of the seafloor.

Static Tsunami Generation Model A model of displacement of the water column above the source region under the assumption that generation occurs instantaneously. Such models can be computed from a *static earthquake model* (q.v.), or the final results of *kinematic* or *dynamic earthquake models* (q.v.).

Tide Gauge A device for measuring the height (rise and fall) of the tide. In particular, an instrument for automatically making a continuous record of tidal height versus time.

Tsunami Amplitude On a time series record of tsunami elevations, such as a tide gauge record, the maximum or minimum between successive downcrossings relative to ambient sea level.

Tsunami Arrival Time Absolute time of initial wave motion of first arrival at an observing station. Reported in *tsunami catalogs* (q.v.).

Tsunami Catalog A catalog of historical tsunami events, describing various parameters of both the tsunami (e.g., arrival time, and size) and the tsunami source (e.g., origin time and location, earthquake magnitude).

Tsunami Coda Phases of tsunami waves after the first arrival. Characterized by random wave motion with an exponentially decaying envelope.

Tsunami Height On a time series record of tsunami elevations, such as a tide gauge record, difference between successive tsunami crest amplitude and trough amplitude.

Tsunami Maximum-Per-Event Amplitude On a time series record of tsunami elevations, such as a tide gauge record, the maximum amplitude of a tsunami event relative to ambient sea level. One possible measure of tsunami size reported in tsunami catalogs.

Tsunami Origin Time Absolute time of tsunami generation. Usually origin time of earthquake when rupture occurs as reported in *tsunami catalogs* (q.v.).

Tsunami Run-up Maximum vertical height above ambient sea level of tsunami at point of farthest inundation, assuming an upward-sloped topographic profile. One possible measure of tsunami size reported in *tsunami catalogs* (q.v.).

Definition of Subject

The devastating consequences of the 2004 Indian Ocean and 2011 Tohoku-Oki tsunamis have led to increased research into many different aspects of the tsunami phenomenon. In this paper, we review research related to the observed complexity and uncertainty associated with tsunami generation, propagation, and occurrence described and analyzed using a variety of stochastic models. In each case, tsunamis generated by earthquakes are primarily considered. Stochastic models are developed from the physical theories that govern tsunami evolution combined with empirical models fitted to seismic and tsunami observations, as well as tsunami catalogs. These stochastic models are key to providing probabilistic forecasts and hazard assessments for tsunamis. The stochastic methods described here are similar to those described for earthquakes (Vere-Jones 2013) and volcanoes (Bebbington 2013) in this Encyclopedia.

Introduction

Tsunamis are generated in the ocean by rapidly displacing the entire water column over a significant area (~ 10 to ~ 100 s of km in length and width). There are three primary submarine geologic phenomena that generate tsunamis: large-magnitude earthquakes, large landslides, and volcanic processes. Earthquakes are by far the most common cause of tsunamis and will be the primary focus of this paper. Generally, earthquakes greater than magnitude (M) 6.5–7 can generate tsunamis if they occur beneath an ocean and if they result in predominantly vertical displacement. However, earthquake rupture does not have to penetrate the seafloor in order to generate vertical (and horizontal) displacement at the seafloor. To simulate the tsunami generation process, elastic deformation models compute the magnitude and pattern of seafloor displacement from earthquake source parameters. For seismogenic tsunamis, the initial wavefield of a tsunami is almost identical to the vertical seafloor displacement field, except in cases of surface rupture and

steep bathymetry where horizontal displacements also contribute to tsunami generation (Satake 2007; Tanioka and Satake 1996). Static tsunami generation models assume that the earthquake occurs instantaneously relative to the phase speed of the tsunami. More recently, dynamic rupture models have been used to more accurately simulate the time-dependent tsunami generation process, using laboratory-derived frictional parameterizations and physically motivated fault stress distributions.

Complexities in the tsunami wavefield during propagation are caused by both spatial variations in slip during the tsunami generation process and by effects such as scattering and reflection. The marked variations in fault slip along the rupture surface are filtered to some extent by the elastic response (Geist and Oglesby 2014) of surrounding material in calculating seafloor displacements and, hence, the initial tsunami wavefield. In the far-field, these variations are smoothed even more by the effects of geometric spreading and dispersion during transoceanic propagation, although additional complexities are introduced by scattering and reflections from geometrically complex coastlines. In the near-field, refractive focusing and amplification during shoaling generally preserves variations derived from the complex source processes.

Stochastic processes (McKane 2016) have been used to reproduce the variability associated with tsunami generation and propagation. Stochastic processes describe physical phenomena, such as Brownian motion, by a set of random variables from a probabilistic point of view (Bartlett 1978; Beichelt and Fatti 2002). Stochastic processes can be either discrete or continuous, commonly as a function of time. A stochastic process can also describe a random field in space, such as been done for describing either static slip (static generation models) or stress applied to a fault (dynamic generation models).

Stochastic models developed from random wave and vibration theory are adapted to describe tsunami observations during propagation. In an early study, Carrier (1970) used stochastically derived bottom topographies to determine that very little of the total tsunami energy is deterred

by scattering in reaching any given target. However, the combination of propagation effects such as scattering with spatial complexity of the source results in a randomly varying tsunami wavefield, similar to what is oceanographically observed in a random sea state from wind waves (Ochi 2005). The random wavefield is observed at recording stations both near and far from the source as time series observations.

Both earthquakes and landslides exhibit some aspects of self-organized criticality (Hergarten 2003; Hergarten and Neugebauer 1998; Werner and Sornette 2013). One consequence of this is scale invariance, in which the sizes of earthquake and landslide tsunami sources tend to follow a Pareto distribution (e.g., Kagan 2002; ten Brink et al. 2006). As a result, tsunami sizes as indicated by maximum-per-event amplitude have been shown to also follow a Pareto distribution at individual recording stations. The size distribution of tsunami sources and tsunamis themselves must be tapered or truncated at some maximum size owing to finiteness of physical dimensions that govern these phenomena, although the point at which sizes deviate from power-law scaling has been difficult to estimate (Geist and Parsons 2014).

Finally, the null hypothesis of how tsunami sources occur in time is that they are randomly distributed following a stationary Poisson process. Tsunami catalog data can be used to test this hypothesis. Relative to a Poisson process, temporal clustering is evident in both the sources and tsunamis observed on recording stations such as tide gauges, with inter-event times described by a gamma distribution (Geist and Parsons 2008, 2011). Clustering can be simulated using stochastic branching processes that have been developed to understand seismicity (e.g., Kagan 2010).

The aforementioned topics are reviewed in the sections below. In the Stochastic Models of Tsunami Generation section, both static and dynamic rupture models are described with respect to tsunami generation. This section is followed by the Stochastic Models of Tsunami Time Series section in which a phenomenological description of tsunami time series during propagation is explained using concepts of random wave theory from both seismology and oceanography.

Whereas the first two sections focus on tsunamis as individual events, the next section, Stochastic Models of Tsunami Occurrence, examines tsunamis as a catalog or an ensemble of events. This section includes a description of the size distribution for tsunamis and their sources, as well as stochastic models of tsunami occurrence in time. The final section, Probabilistic Analysis of Tsunami Hazards, synthesizes concepts offered in this paper toward producing tsunami hazard assessment tools for engineering and risk-evaluation purposes.

Stochastic Models of Tsunami Generation

Stochastic models of earthquakes have been developed to account for the observed complexity of slip and stress drop distribution. Static stochastic source models are considered first. Using static source models for tsunami generation is based on the assumption that the rupture process time is small relative to the time it takes for tsunami waves to move away from the source region. While this assumption may be adequate for many earthquakes, for long earthquake ruptures the slip history does have an effect on the tsunami wavefield (e.g., Satake et al. 2013). Dynamic tsunami generation models are also discussed below to address these ruptures and to provide a self-consistent physical model between stresses resolved along the fault and the resulting slip distribution.

Static Slip Models for Earthquakes

Stochastic models of earthquake rupture complexity have been developed over the last several decades, most notably starting with Andrews (1980) in his examination of static slip. Static stochastic models for tsunami generation by earthquakes were introduced by Geist (2002) as an alternative to uniform-slip dislocation models that had been previously used to simulate tsunami generation. This model of tsunami generation combines Andrews' (1980) stochastic static models for slip along a fault with the point-source elastic dislocation expressions of Okada (1985) to

compute seafloor displacement. Refinements of stochastic static slip model are detailed below, followed by a description of recent tsunami generation models based on dynamic earthquake rupture under stochastic stress conditions on a fault system.

The stochastic static slip model of Andrews (1980) is based on the idea that the final slip distribution after an earthquake rupture is self-affine. In contrast to isotropic scaling associated with a self-similar fractal, self-affine fractals (Turcotte 2013) are nonisotropic (Turcotte 1992). The physical origin of heterogeneous slip is likely caused by complex dynamic changes in the stress field during earthquake rupture and/or fault roughness or other fault zone heterogeneities. Fault roughness has been shown to be self-affine over many orders of magnitude (Candela et al. 2012).

The slip spectrum in the radial wavenumber domain k is given by (after Herrero and Bernard 1994)

$$D(k) = C \frac{\Delta\sigma}{\mu} \frac{L}{k^\gamma} k > k_c, \quad (1)$$

where $\Delta\sigma$ is the mean stress drop, μ the shear modulus, L the characteristic rupture dimension, and C a constant. Below the corner wavenumber k_c , the spectrum is assumed to be constant and proportional to the mean slip for the entire rupture (Herrero and Bernard 1994). In this more general form of the slip spectrum, the spectral decay exponent γ is variable and linked to the far-field seismic displacement spectrum (Hisada 2000, 2001; Lavallée 2008), whereas in Andrews (1980) and Herrero and Bernard (1994) $\gamma = 2$. This spectral decay exponent is related to Hurst exponent (H) for fractional Brownian motion (Peitgen et al. 1992; Tsai 1997). The description of the stochastic slip model in the spectral domain is related to the autocorrelation function according to the Wiener-Khinchine theorem (Andrews 1980). Values of H such that $0.5 < H < 1$ indicate long-range positive autocorrelation.

To produce stochastic slip from this spectral model, phase is randomized, producing different realizations of slip distribution (Andrews 1980; Herrero and Bernard 1994). Equivalently

$$D(x) = D_0 F^{-1}[R_s(k)D(k)], \quad (2)$$

Where $F^{-1}[\cdot]$ is the inverse Fourier transform, $R_s(k)$ is the Fourier transform of a random function $R(x)$ that provides the stochastic kernel for the slip distribution, and D_0 is a constant (Lavallée 2008; Lavallée et al. 2006; Liu-Zeng et al. 2005). Figure 1 shows three slip distributions using different spectral decay exponents, but keeping the same phase information so that regions of high/low slip occur at the same spatial position on the fault.

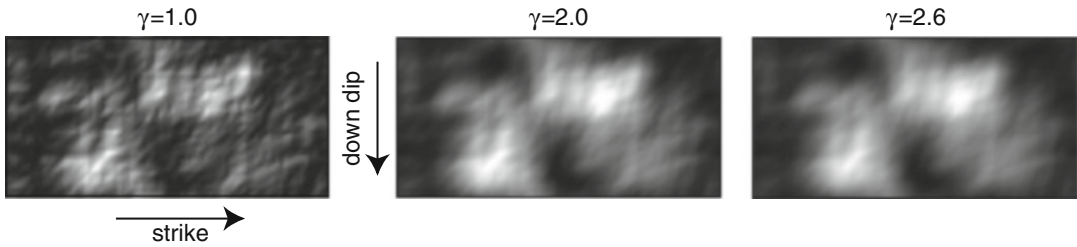
Mai and Beroza (2002), using a von Kármán power spectrum, provide an alternative static stochastic slip model for earthquakes. In this case, different correlation lengths are used in the dip and strike directions, where the power spectrum is given by

$$D(k) = \frac{4\pi H}{K_H(0)} \frac{a_x a_z}{(1 + k^2)^{H+1}}, \quad (3)$$

H is the Hurst number and K_H is the modified Bessel function of order H . Mai and Beroza (2002) indicate that the correlation lengths a_x and a_z scale with the moment magnitude of the earthquake. As an important refinement for large-scale ruptures, LeVeque et al. (2016) have developed a stochastic slip model for general fault geometries, such as non-planar subduction zones.

Most recently, Lavallée et al. (2006, 2011) indicate that slip determined from the inversion of seismic waveforms is more variable than the Andrews (1980) model specifies. Previous slip models primarily used a Gaussian distribution for $R(x)$. Lavallée et al. (2006) indicate that the heavy tail Lévy α -stable distribution family produces a closer match to slip inversions than the Gaussian distribution. Davies et al. (2015) perform a comprehensive comparison of the aforementioned stochastic slip models and emphasize the importance of testing a particular methodology with respect to tsunami inundation calculations.

Using these stochastic slip models for tsunami generation involves calculating static elastic displacements for a discretized representation of slip, since the slip functions are not analytically



Tsunamis: Stochastic Models of Occurrence and Generation Mechanisms, Fig. 1 Stochastic, self-affine slip within a rectangular rupture using the Herrero and Bernard (1994) slip model. Slip amplitude given by gray scale

smooth. Surface displacements are calculated using the point-source dislocation expressions such as those derived by Okada (1985) for a uniform half-space. The displacement from each sub-fault element can be linearly superimposed to determine the total displacement field (Satake and Kanamori 1991). See Geist and Oglesby (2014) for more details on these elastic deformation models and for elastically heterogeneous models. Static surface displacements calculated from the elastic deformation models provide initial conditions to tsunami propagation calculations. Early studies incorporating stochastic slip used the hydrostatic shallow-water wave equations. Løvholt et al. (2012) demonstrated that including non-hydrostatic terms in the wave equations (i.e., dispersion) reduces the variability in tsunami run-up from stochastic slip.

There are a number of regional studies that employ stochastic slip for deterministic hazard studies. For example, Geist (2002) and González et al. (2009) use the Herrero and Bernard (1994) slip model to determine the variation of nearshore tsunami amplitudes and run-up along the Mexican and Cascadia subduction zones, respectively, using a finite-difference approximation to the shallow-water wave equations. McClosky et al. (2008) use a fractal slip model consistent with fractal dimension observations by Mai and Beroza (2002) as input to an elastic finite-element model to assess the tsunami hazard in the Indian Ocean. Ruiz et al. (2015) calculate near-field run-up from subduction earthquakes in northern Chile and indicate that stochastic slip produces significantly higher run-up compared to a uniform-slip model, particularly for stochastic realizations that involve high slip near the trench. A number of studies

(e.g., Fukutani et al. 2015 and Goda et al. 2014, 2015b) have examined the role of variable slip on tsunami severity for an event like the devastating 2011 Tohoku-Oki earthquake. Consistent with the results of Ruiz et al. (2015), high slip near the trench, such as that occurred with the 2011 event, results in significantly higher run-up and inundation.

Finally, a series of studies examine the expected variation in near-field tsunami edge waves, using stochastic slip for continental subduction earthquakes. Edge waves are waves fixed by refraction at the coast, propagating parallel to the shoreline, and often result in the largest tsunami amplitudes for sites at oblique ray paths from the source. Geist (2012a) combines initial conditions derived from the Lavallée et al. (2006) slip model with analytical edge wave theory where the shore-parallel wavefield is decomposed into Fourier components (Carrier 1995; Kirby et al. 1998; Mei et al. 2005). Geist (2012a) finds significant variation in the edge wavefield imparted by stochastic slip compared to simpler characterizations of the earthquake source, such as for crack or dislocation models. Geist (2016, 2018) further examines nonlinear resonance associated with edge wave modes forced by a stochastic source. Results indicate that there is a direct connection between the probability distribution of the random function $R_s(k)$ used in the Lavallée et al. (2006) stochastic model and which edge wave modes participate in resonance.

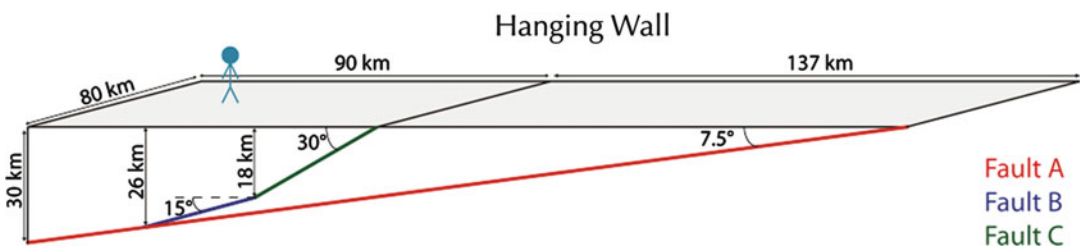
Dynamic Rupture Models for Earthquakes

Whereas stochastic slip models can efficiently produce a suite of likely rupture and slip scenarios for tsunami generation, they are not

fundamentally linked to the physics of the earthquake process. As a consequence, there is no guarantee that any such slip model is consistent with a physically justifiable stress drop pattern or friction law on the fault. Thus, such slip models may include rather unrealistic and unlikely slip patterns. Dynamic earthquake faulting models (Andrews 1976; Das 1981; Day 1982; Harris et al. 1991; Ma and Beroza 2008; Mikumo and Miyatake 1993; Oglesby et al. 1998; Olsen et al. 1995) can help to solve these problems. Unlike kinematic slip and/or dislocation models, dynamic models of faulting start from a (potentially stochastic) initial stress field on the fault surface, as well as physically and observationally based estimates of frictional and material parameters on and near the fault system. The method then solves the coupled equations of motion and friction on a grid in space and time after a prescribed instigation location for the earthquake. Rather than being assumed a priori, the evolution of rupture and the final slip pattern, as well as the time-dependent ground deformation, are calculated results of the numerical model. Of course, significant uncertainties still exist, particularly in the initial stress field and the frictional properties. However, the faulting process and ground motion are more likely to be consistent with both the fault geometry and our understanding of the basic physics of the earthquake process: properties such as dynamic interactions between different fault segments, the effects of the free surface on rupture propagation, and the ability of rupture to penetrate low-stress regions are automatically calculated rather than being assumed.

By producing many dynamic models, using a variety of assumptions about the physical parameters, one may bracket the physical possibilities in a given fault system and also obtain an estimate of which faulting behaviors are more likely across a wide range of assumptions. For example, if dynamic rupture penetrates a known zone of low seismic coupling under almost all scenarios, one may give that behavior higher weight in the tsunami hazard calculation; kinematic modeling by itself does not give such insight. Finally, dynamic modeling can give crucial physical insight into the physical source of the rupture and ground motion behavior we see in past earthquakes, opening up the possibility of estimating such effects in future events.

Dynamic rupture models have been used to explore the physics of branched faults for some time, including the pertinent case of megathrust/splay fault systems (e.g., Aochi et al. 2000, DeDontney et al. 2012, Duan and Oglesby 2007, Kame et al. 2003, Tamura and Ide 2011, and Wendt et al. 2009). A description and example of the use of dynamic models for tsunami generation is given in Wendt et al. (2009) as well as Geist and Oglesby (2014). Herein we show how dynamic modeling may be used to produce stochastic models of tsunami generation and propagation. The method is essentially the same as Wendt et al. (2009), except that we use a stochastic distribution of shear stress on the fault system and run multiple realizations of the random variables to produce a suite of earthquake and tsunami models. Figure 2 shows a cartoon of the model fault geometry, which is motivated by a possible



Tsunamis: Stochastic Models of Occurrence and Generation Mechanisms, Fig. 2 Fault geometry of Wendt et al. (2009). The splay fault (faults B and C) branches off

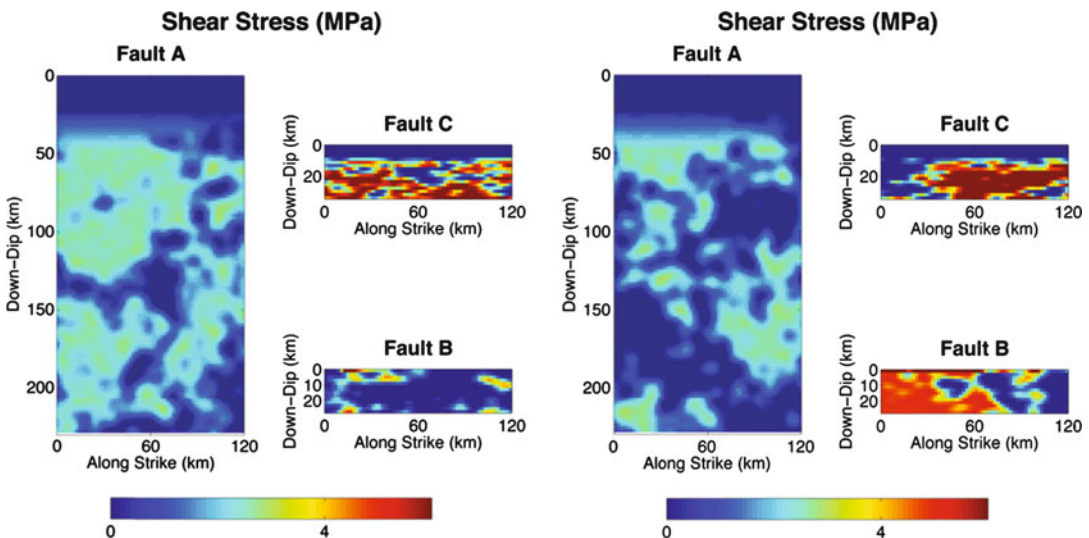
the plate boundary thrust (fault A) close to shore and at a significantly steeper dip than the plate boundary thrust

splay fault (a subsidiary fault that branches from the main fault) on the Nankai Trough subduction zone.

Rather than assume a self-affine slip distribution on this geometrically complex fault system, we use a slightly modified version of the method of Andrews and Barall (2011) to assign stochastic patterns of shear stress to each fault segment. This method, derived for general earthquake modeling, assumes that the spatial variation in initial (prior to the earthquake) shear stress is a self-similar random function (i.e., a power-law function with a Hurst exponent of zero) with a low-spatial-wavelength cut-off of 6 km. The resulting randomized stress distributions consist of high-stress asperities surrounded by low-stress background areas, with stress amplitude constrained to be below the threshold that would cause the fault to slip prior to the earthquake initiation. Stress patterns on nearby fault segments are not constrained to be similar; such a method is physically desirable and under development, but beyond the scope of this initial work. Beyond a depth of 6 km, the average normal stress is constant (4.77 MPa on segment A, 9.04 MPa on segment B, and 12.38 MPa on segment C), resulting in no imposed depth dependence on stress drop beyond

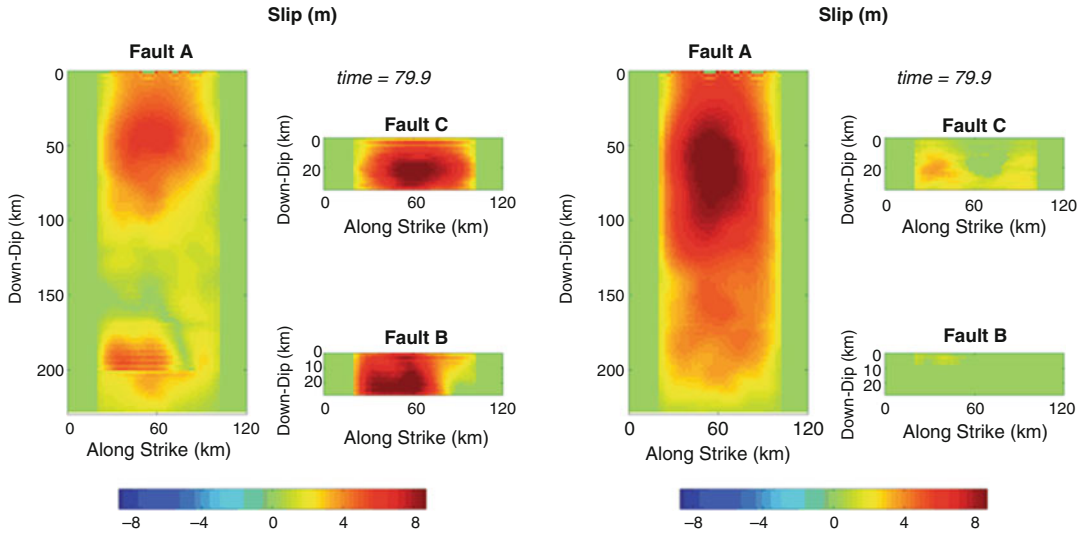
the random distribution. Between 6 and 3 km depth, the normal stress is linearly scaled down to 1/10 of its ambient value, and the shear stress is scaled down to zero; these stresses remain at those values up to the surface. This process is to mimic both the decrease of the lithostatic load and the rate-strengthening frictional behavior believed to exist near the surface of many subduction zones. An example of two such stress distributions is given in Fig. 3.

With these stress distributions, the dynamic finite-element method (Oglesby 1999; Whirley and Engelmann 1993) can be used to perform spontaneous dynamic rupture models of potential earthquakes on this fault system. As in Wendt et al. (2009), we use linear time-weakening friction with a weakening time of 0.5 s; this method results in an effective slip-weakening distance that is spatially dependent and does allow us to better resolve the weakening process throughout the calculation. Continuing our example, the fault slip distributions resulting from the stress distributions in Fig. 3 are shown in Fig. 4. Note that the left example has greater slip on the plate boundary thrust fault (Fault A) and less slip on the splay fault segments (Faults B and C) than the right example. By examining the stress patterns, we



Tsunamis: Stochastic Models of Occurrence and Generation Mechanisms, Fig. 3 Two randomization realizations of shear stress on the fault system in Fig. 2. The

white line indicates the location of the intersection of the megathrust and lower splay fault



Tsunamis: Stochastic Models of Occurrence and Generation Mechanisms, Fig. 4 Two slip distributions corresponding to the randomized stress distributions in

Fig. 3. The white line indicates the location of the intersection of the megathrust and lower splay fault

can see the source of these differences in slip patterns. In particular, note that in the left example, there is very little shear stress on the splay fault near the intersection of the megathrust. This lack of stress inhibits rupture propagation to the splay fault. The important point in this example is that different stochastic stress distributions can lead to an activation of the splay fault or not. We also note that the two seismic moments are not equal; the one on the left is 1.8×10^{21} Nm (Mw 8.14), while the one on the right is 1.5×10^{21} Nm (Mw 8.08).

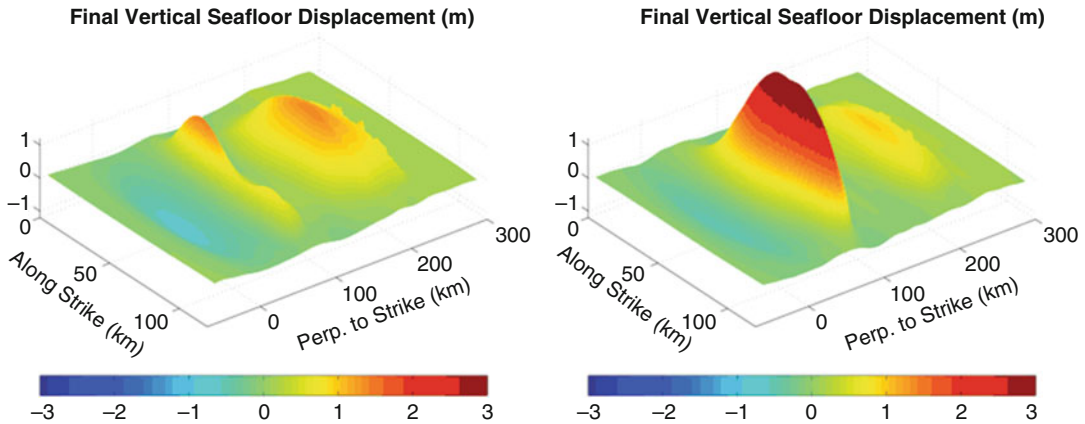
In turn, these two very different slip distributions lead to very different vertical surface deformation patterns, as shown in Fig. 5. In particular, the right panel that corresponds to the model with significant splay fault slip has a much higher ground deformation around the splay fault. This result will have important implications for tsunami generation, as we shall see in Probabilistic Analysis of Tsunami Hazards section.

Recently, Andrews and Ma (2016) specify a stochastic initial stress distribution that ties to a target rupture length and moment magnitude. In that study, the presumed stress drop tapers at the edges of the fault so that the rupture stops in a progressive fashion. A presumed mean stress drop

is chosen to produce the desired (approximate) moment magnitude, while the resultant ground motion spectrum agrees with some empirical ground motion prediction equations. However, we note that in general, one cannot prescribe an exact moment magnitude for dynamic rupture models; rather, it is a calculated result from the physics of the model. The Andrews and Ma (2016) procedure involves going from 2-D wavenumber space to a 2-D spatial domain via the inverse Fourier transform. The wavenumber space follows a power-law distribution as shown in Fig. 6. Similar to Andrews and Ma (2016), the initial shear stress function (τ_0) on the fault can be parameterized by

$$\tau_0 = [\mu_k + \gamma w(x, d)] C(d) \sigma_0, \quad (4)$$

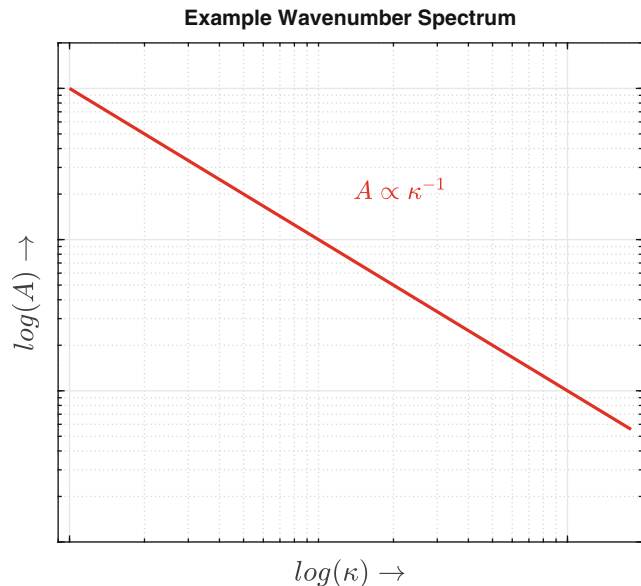
where μ_k is the kinetic (i.e., dynamic) friction coefficient, γ scales w which is a random self-similar 2-D function that depends on along-strike and downdip coordinates x and d , respectively, $C(d)$ is a downdip-conditioning function, and σ_0 is some constant initial normal stress. However, as seen in Andrews and Barall (2011) and Andrews and Ma (2016), a variable normal stress function can be used to reduce stress drop near the



Tsunamis: Stochastic Models of Occurrence and Generation Mechanisms, Fig. 5 Two vertical seafloor displacement distributions corresponding to the randomized

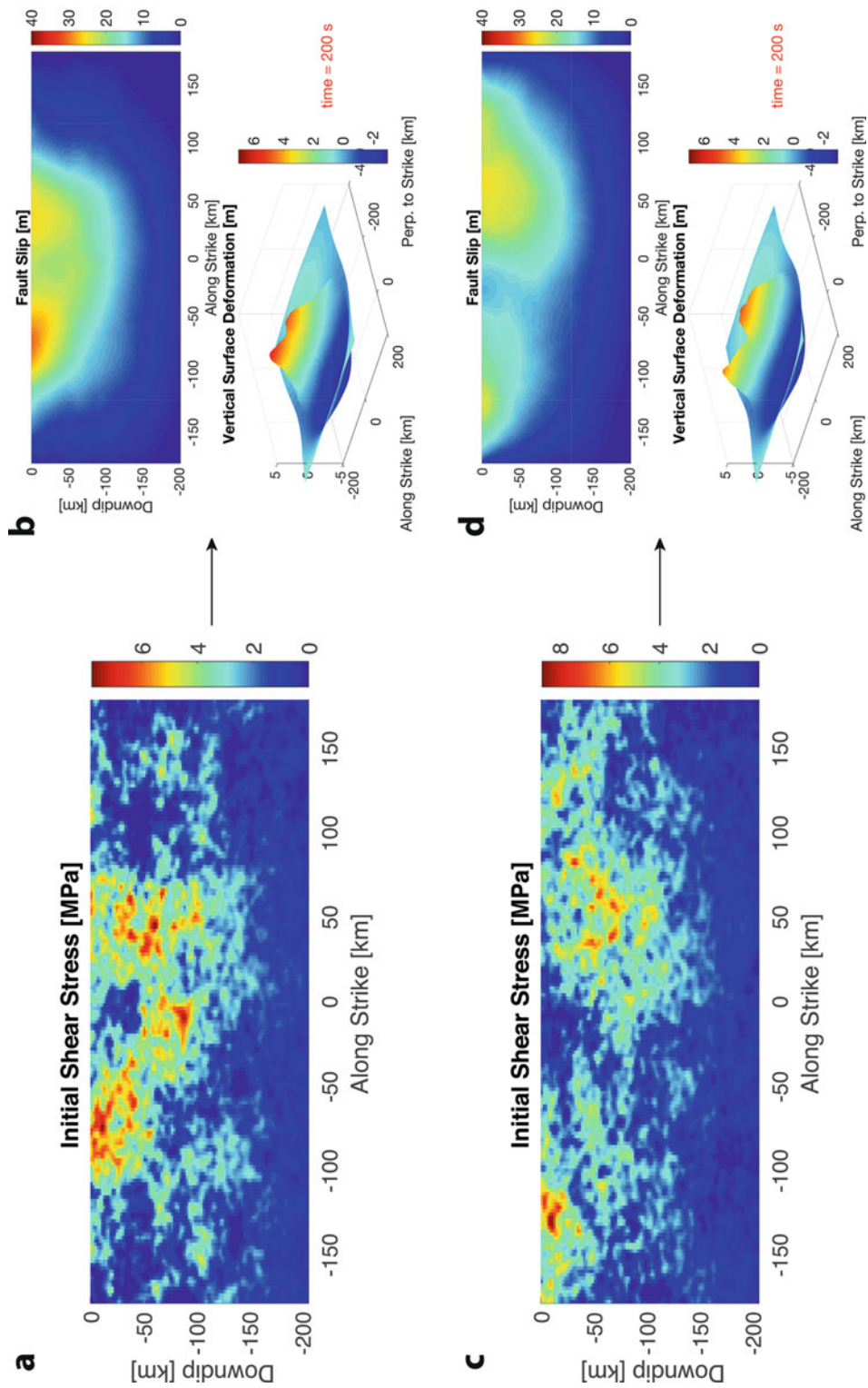
stress distributions in Fig. 3. The shoreline is in the negative direction along the perpendicular-to-strike axis

Tsunamis: Stochastic Models of Occurrence and Generation Mechanisms, Fig. 6 An example log-log wavenumber (κ) spectrum with exponent -1 and amplitude (A) (cf. Andrews and Ma 2016)



free surface (i.e., the seafloor). In using the method above, a target earthquake magnitude can be produced with several different initial stochastic shear stress distributions. For example, we incorporate this method and use the dynamic finite-element method (Ryan et al. 2015) to model a megathrust fault rupture and the resulting seafloor deformation (Fig. 7) along the Alaskan-Aleutian subduction zone. Therefore, we generate a 2-D set of complex random amplitudes that conform to a κ^{-1} falloff (see Fig. 6) and modify

those amplitudes at low wavenumber modes so that the mean value is approximately zero and so that the half-wavelength modes that correspond to the along-strike length of the fault have the largest influence on rupture propagation length. This shear stress function (inverse Fourier transform from the wavenumber domain) is normalized by the low wavenumber modes so that the overall shear stress envelope is determined by the low wavenumber modes. The shear stress tends to zero at the edges of the fault with the exception



Tsunamis: Stochastic Models of Occurrence and Generation Mechanisms, Fig. 7 Two example initial random shear stress distributions in the left column, with their resultant fault slip and vertical seafloor deformation patterns in the right column. Both modeled earthquakes have a moment magnitude $M_w = 9$

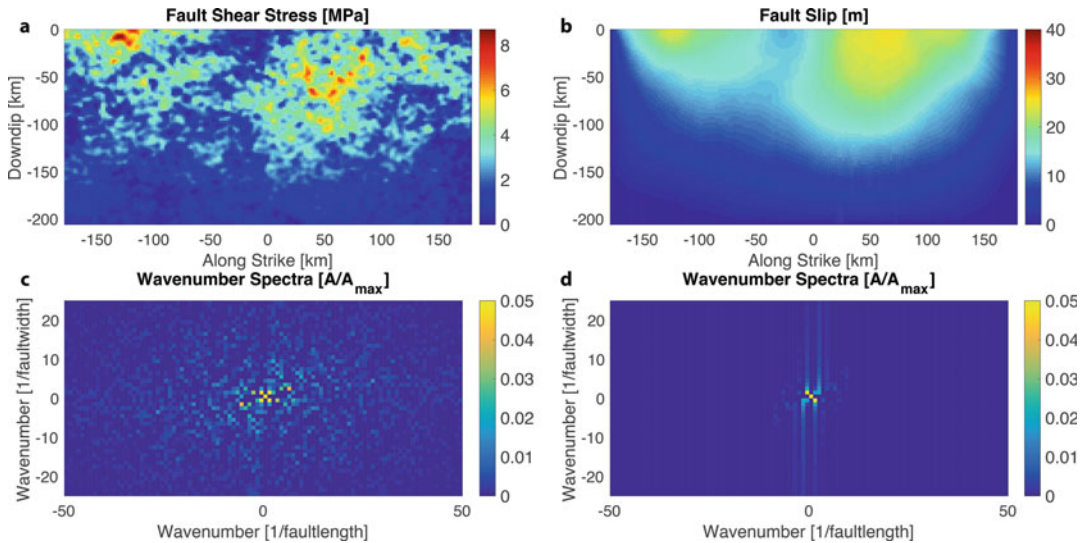
of the free surface. Figure 7 shows that, given the same resultant moment magnitude $M_w \approx 9$, very different slip distributions (panels b and d, Fig. 7) can result from initial random shear stress distributions (panels a and c, Fig. 7).

The reader may notice that the resultant slip distributions in Fig. 7 are much “smoother” than their corresponding initial shear stress distributions. However, the slip distributions do show heterogeneous patterns that qualitatively resemble the corresponding initial shear stress distributions. Using Fourier analysis on Fig. 7c, d, we plot example initial shear stress and fault slip spectra in Fig. 8. Note that the wavenumber spectra are normalized by the maximum amplitudes and that the units for wavenumbers are $1/\text{faultlength}$ where faultlength is the length (km, along strike) of the fault and $1/\text{faultwidth}$ where faultwidth is the width (km, downdip) of the fault. The wavenumber spectrum for the slip distribution is localized to relatively lower wavenumbers (i.e., larger wavelength components) than the corresponding spectrum of initial shear stress. Thus, we find that relatively rougher shear pre-stress distributions produce relatively smoother, but still highly heterogeneous, slip distributions.

Stochastic Models of Tsunami Time Series

In this section, we demonstrate how tsunami time series observations recorded on both deep-ocean pressure sensors and coastal stations are aided by using stochastic models. The first arrival on unobstructed propagation paths preserves the waveform characteristics from tsunami generation, excluding any shock waves (i.e., steeply sloped waves) that might develop over long travel distances. Later arrivals are influenced by a variety of effects, including dispersion, scattering, refraction, trapped waves, reflected arrivals, and resonance (González and Kulikov 1993; Koshimura et al. 2008; Kowalik et al. 2008; Mofjeld et al. 2001; Saito and Furumura 2009; Satake et al. 1992). Perhaps surprisingly, both coastal and deep-water time series records of tsunamis can be described by a similar stochastic model as described below.

The tsunami coda (i.e., phases later than the first cycle) can be described using concepts from both seismic ground motion and oceanographic observations of a random sea. The tsunami coda is a random wavefield modified by an exponentially



Tsunamis: Stochastic Models of Occurrence and Generation Mechanisms, Fig. 8 Example initial random shear stress distribution and corresponding wavenumber spectrum in the left column, with the resultant fault slip distribution and corresponding wavenumber spectrum in

the right column. The faultlength and faultwidth parameters are the length (km, along strike) of the fault and the width (km, downdip) of the fault, respectively, and each wavenumber spectrum is normalized by the maximum amplitude ($A_{\max}$) for visual clarity

decaying envelope, similar to the seismic coda from earthquakes (Aki and Chouet 1975). The stochastic model of water-level elevations (η) within the coda is that of a time-varying Gaussian distribution (Takahara and Yomogida 1992). The stationary Gaussian distribution is also the canonical model for deep-water waves in a fully developed random sea (Longuet-Higgins 1952), though Gaussian transformations (Rychlik et al. 1997) and second-order statistical models have been proposed that more accurately reflect observed wind-generated waves in the ocean (e.g., Hogben 1990, Huang et al. 1990, Jha and Winterstein 2000, Muraleedharan et al. 2007, and Prevosto et al. 2000). These latter models may also provide a more refined description of the tsunami wavefield in the future. The envelope of tsunami amplitudes is assumed to follow an exponential decay of the form (Mofjeld et al. 2000)

$$A_c = A_0 \sigma \exp[-(t - t_0)/\tau], \quad (5)$$

where A_0 is a constant coefficient, σ is the initial standard deviation of A_c , and τ is the e -folding decay constant (the time an exponentially decaying process decreases by a factor e , analogous to Q for the seismic coda). The e -folding decay constant for the tsunami coda is 22.0 ± 0.7 h as measured by van Dorn (1984) for 28 tsunami sources in the Pacific. Rabinovich et al. (2011), however, measured a wider range of decay constants (13–45 h) for the 2004 Indian Ocean tsunami throughout the world's oceans.

Statistical tests can be performed to determine whether the coastal tsunami coda can in fact be described as a time-varying Gaussian distribution (Geist 2009). The data we used for this study are time series observations of the 2006 and 2007 Kuril Islands tsunamis. The earthquake that generated the 2006 tsunami was a $M_w = 8.3$ plate boundary thrust earthquake that triggered the 2007 $M_w = 8.1$ outer-rise, normal faulting earthquake (Lay et al. 2009). These two events present a unique opportunity to determine whether earthquake mechanism affects the statistical characteristics of the tsunami on the same observation stations and for similar source locations and

magnitudes. First, data from the 2006 Kuril Islands tsunami for both a continental station (Crescent City) and an island station (Midway Island) are examined (Fig. 9a, b, respectively).

Estimation of the coda envelope indicates that the e -folding time for Midway Island is 22 h, consistent with observations by van Dorn (1984). The empirical density distribution of water-level elevations detided and corrected for the exponential envelope decay is shown in Fig. 10. According to the Shapiro-Wilk test, the Gaussian null hypothesis cannot be rejected for either the Crescent City or the Midway Island data at the 95% confidence level.

The distribution of wave heights (H) corresponding to the aforementioned Gaussian distribution of water elevations is a Rayleigh distribution (Rayleigh 1880). Longuet-Higgins (1952) derived this correspondence for the case of a narrowband of frequencies for ocean waves. The Rayleigh density distribution is typically given in terms of the crest-to-trough wave height

$$f(H) = \frac{2H}{H_{\text{rms}}^2} \exp\left(-\frac{H^2}{H_{\text{rms}}^2}\right), \quad (6)$$

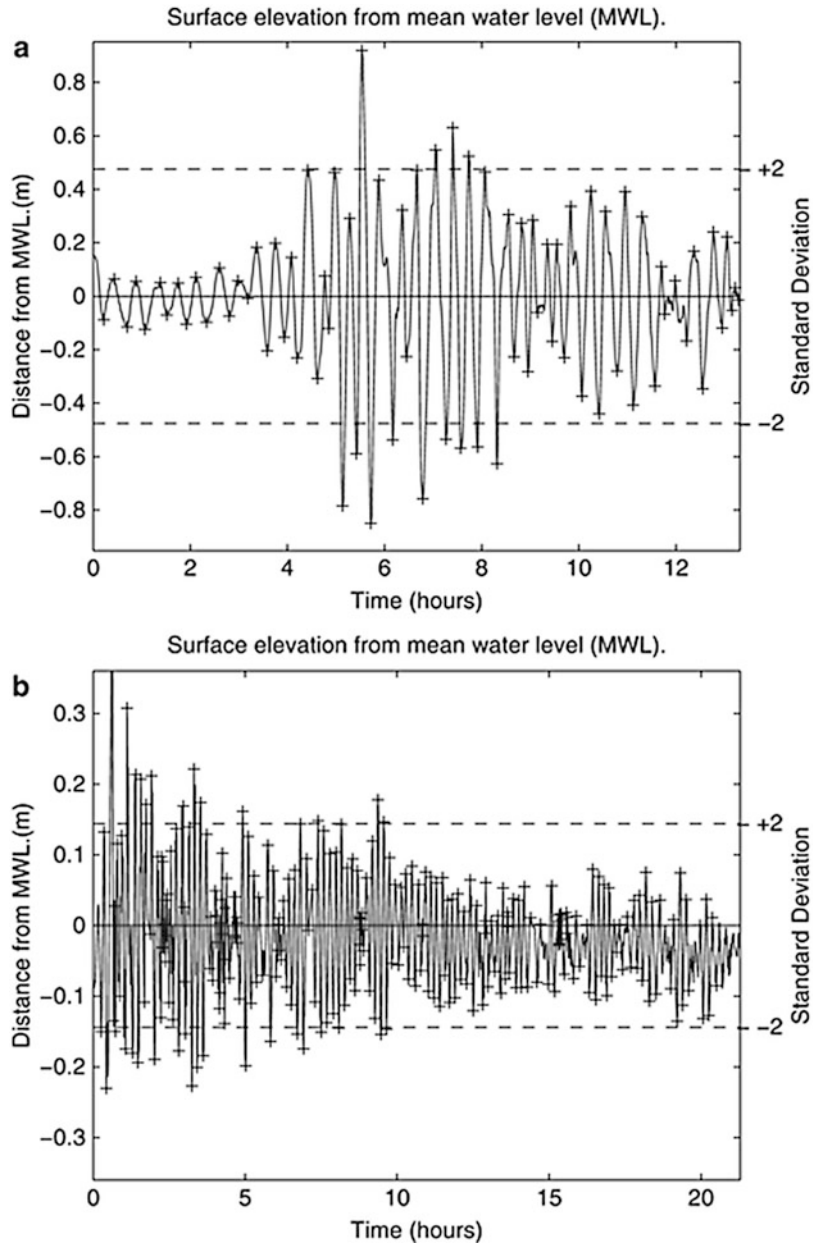
where H_{rms} is the root-mean-square wave height. For amplitudes, the distribution can be written as (Massel 1996)

$$f(A) = \frac{A}{\sigma_\eta^2} \exp\left(-\frac{A^2}{2\sigma_\eta^2}\right), \quad (7)$$

where σ_η^2 is the variance of the water-level elevation time series (e.g., Fig. 9). For a stochastic time series, definitions of wave amplitude and wave height are not necessarily straightforward. Amplitude can be defined as $A = \frac{1}{2}H$ (Longuet-Higgins 1952), or the crest (trough) amplitude A_c (A_r) can be defined as the global maximum (minimum) between successive downcrossings (Rychlik and Leadbetter 1997) as shown in Fig. 11. In this case, $H = A_c - A_r$. Shown in Fig. 12 is the empirical cumulative distribution of the wave heights shown in Fig. 11 in comparison to the corresponding Rayleigh distribution. Including the direct arrival wave height can result in a poorer

Tsunamis: Stochastic Models of Occurrence and Generation Mechanisms, Fig. 9

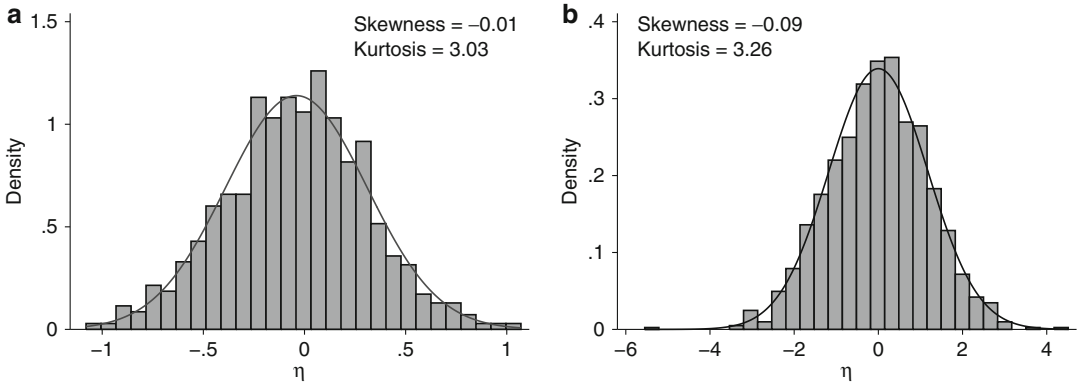
Time series of tide gauge measurements of the 15 November 2006 Kuril tsunami: (a) Crescent City (6293 km from epicenter), (b) Midway Island (3271 km from epicenter). Location of crest and trough amplitudes indicated by '+'s



fit when it is much greater than the tsunami coda wave heights (e.g., Fig. 12d).

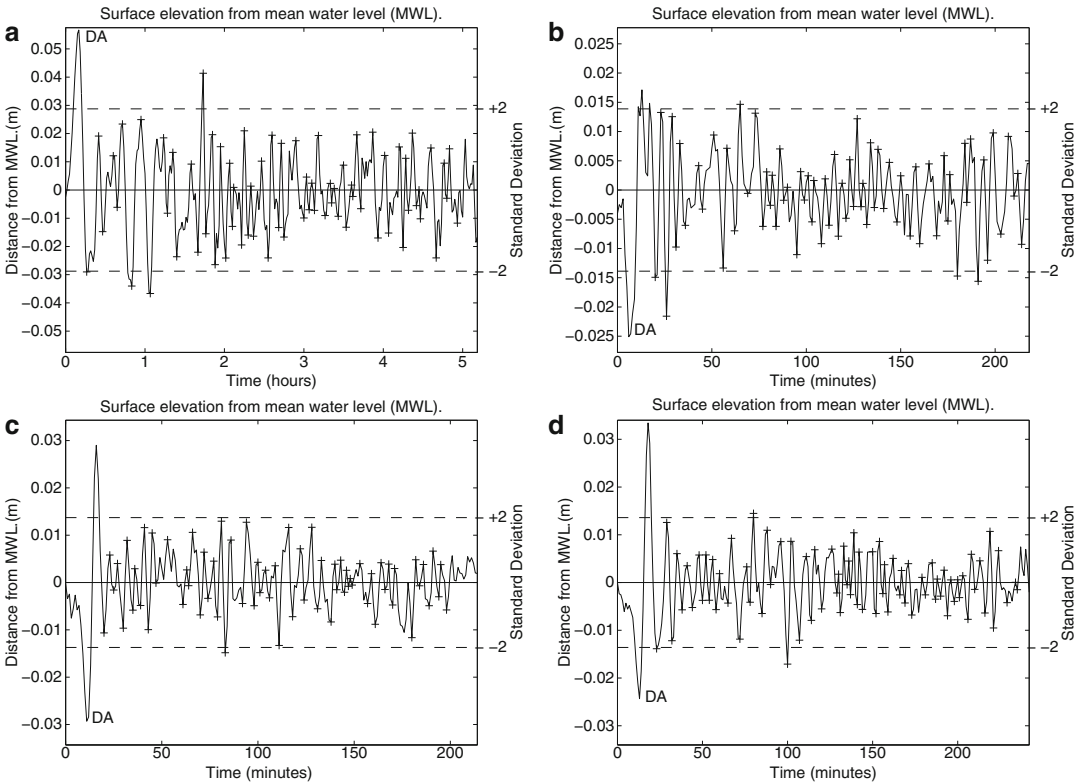
Understanding the stochastic nature and envelope of tsunami wave heights is key to determining when tsunami warning and advisories may be lifted. For example, Mofjeld et al. (2000) develop

a forecast methodology for Pacific-wide tsunami waves using the envelope of the stochastic tsunami coda combined with background water levels from tides and lower-frequency oscillations. The amplitude of the envelope approximately scales with the moment magnitude of the



Tsunamis: Stochastic Models of Occurrence and Generation Mechanisms, Fig. 10 Empirical density distribution of tsunami coda waves from data plotted in Fig. 6:

(a) Crescent City, (b) Midway Island. Line indicates Gaussian distribution model



Tsunamis: Stochastic Models of Occurrence and Generation Mechanisms, Fig. 11 Tsunami time series from Deep-Ocean Assessment and Reporting of Tsunamis (DART) bottom-pressure stations operated by NOAA, showing the direct arrival (DA) and the ensuing tsunami

coda. Location of crest and trough amplitudes indicated by + symbol. (a) 15 November 2006 Kuril tsunami, station 46413. (b) 13 January 2007 Kuril tsunami, station 21413. (c) 13 January 2007 Kuril tsunami, station 21414. (d) 13 January 2007 Kuril tsunami, station 46413

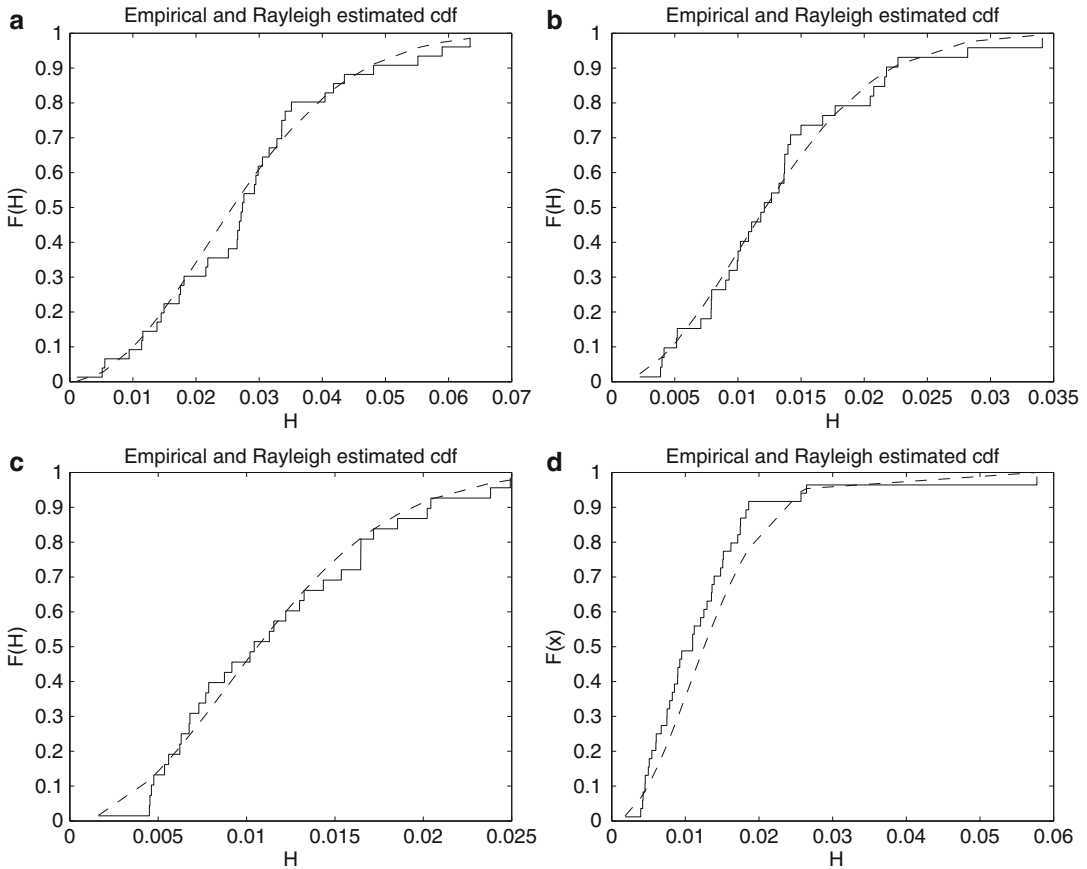


Fig. 8. Dashed line indicates Rayleigh cumulative distribution model

Tsunamis: Stochastic Models of Occurrence and Generation Mechanisms, Fig. 12 Empirical cumulative distribution (solid line) of wave height (H) from data shown in

causative earthquake (Geist 2012b). The forecasting method is tested by Mofjeld et al. (2000) against six instrumentally recorded tsunamis in the Pacific. Tsunami coda forecasts are generally accurate 4 h after local tsunami onset time and complement real-time, deterministic modeling forecasts for earlier waves (Titov et al. 2005).

Stochastic Models of Tsunami Occurrence

Because earthquakes are the primary generator of tsunamis and often provide the trigger for other tsunami sources such as submarine landslides, various empirical relations derived from the

analysis of seismicity can be readily adapted for understanding tsunami occurrence, which is explored in this section. Perhaps the most important of these relations is the distribution of earthquake sizes as directly related to the distribution of tsunami sizes. Other empirical relations, such as the Omori-Utsu relation of aftershock decay, are also important in developing temporal occurrence models for tsunamis.

Both tsunami sources and tsunamis themselves are viewed as point processes in which each point represents the time and location of an event. Because tsunamis can propagate to extremely far distances from the source and can persist for long times (as indicated by the e -folding time in the previous section), at first glance it seems difficult

to determine how point process theory can be applied to analyze tsunamis. However, the overall mean return time for tsunamis, either globally or at a particular recording station is long relative to the event duration. Even for tsunamis occurring close in time relative to the e -folding time, it still is possible to distinguish individual events at individual recording stations. Thus, a point process description is applicable to the study of tsunami occurrence as it is for seismicity (Corral 2009).

Historical Data

Historical data contained in tsunami catalogs provides the foundation for developing and testing stochastic models of tsunami occurrence. This data includes both source parameters and measurements of tsunamis at various locations. Tsunami size (A) is used in general terms, encompassing both maximum-per-event time series amplitude and maximum-per-event run-up measurements. Two separate but cross-referenced tsunami databases are provided at the US National Centers for Environmental Information (NCEI, formerly NGDC): one that is keyed on the source and one keyed on tsunami measurements at a particular coastal location or within a particular region. Most important for developing tsunami occurrence models, the source parameters include the origin time and the source potency, i.e., magnitude for earthquake sources. There are a variety of tsunami measurements, including arrival times, tide gauge and deep-ocean bottom-pressure sensor amplitudes, surveyed run-up and overland flow height measurements, and eyewitness observations.

Distribution and Stochastic Simulation of Tsunami Sizes

Burroughs and Tebbens (2005) originally discovered that tsunami size listed in tsunami catalogs, as measured by maximum-per-event amplitude on individual tide gauge stations, follows a power-law relation, much like the Gutenberg-Richter frequency-size relation for earthquakes (Gutenberg and Richter 1944; Ishimoto and Iida 1939). These relations include a rate of occurrence at some threshold magnitude (logarithm of seismic moment or tsunami size), termed the “ a value” and the slope of the relation when

magnitude is plotted on a linear scale, termed the “ b value.” As a proper probability distribution, the rate is removed, and earthquake and tsunami sizes (A) follow a Pareto distribution above some threshold (A_t) with a density function given by

$$\varphi(M) = \beta A_t^\beta A^{-1-\beta} \text{ for } A_t \leq A, \quad (8)$$

where the Pareto exponent $\beta = \frac{2}{3}b$.

Both tsunamis measured at a particular site and the sources that cause them must be size limited, although the maximum size has been exceedingly difficult to determine (Geist and Parsons 2014). The finite-size effect results in a modified Pareto distribution, in which the tail is either truncated or tapered (Kagan 2002). The truncated form of the Pareto distribution, consistent with the Burroughs and Tebbens (2005) study, is given by

$$\varphi(A) = \frac{\beta A_t^\beta A_x^\beta}{(A_x^\beta - A_t^\beta) A^{1+\beta}} \text{ for } A_t \leq A \leq A_x \quad (9)$$

where $\varphi(A)$ is the density distribution, A_t is the threshold size, and A_x is the truncation parameter. The tapered Pareto distribution, also known as the Kagan distribution (Vere-Jones et al. 2001), with a density function given by

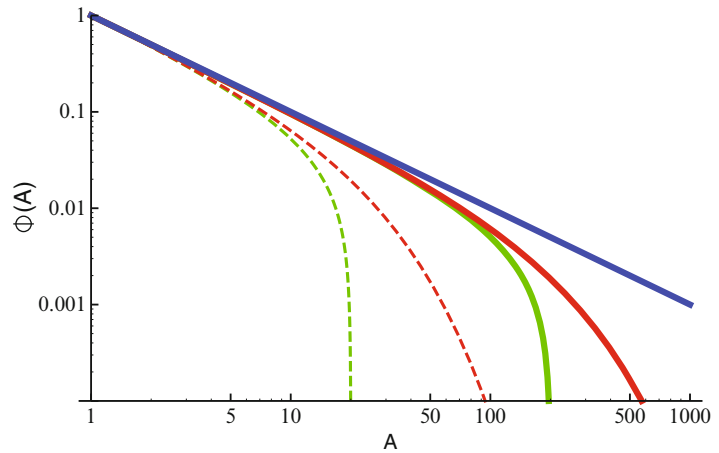
$$\varphi(A) = \left(\frac{\beta}{A} + \frac{1}{A_c}\right) \left(\frac{A_t}{A}\right)^\beta \exp\left(-\frac{A_t - A}{A_c}\right), \text{ for } A_t \leq A \quad (10)$$

has been used in the past (Geist and Parsons 2006; Geist et al. 2009; Parsons and Geist 2009), where A_c is the corner size parameter of the distribution. The gamma-modified Pareto distribution (Kagan 1991) is similar to the tapered Pareto distribution and consistent with finite total seismic energy release (Greenhough and Main 2008; Main 2000; Sornette and Sornette 1999):

$$\varphi(A) = C\beta \frac{A_t^\beta}{A^{1+\beta}} \exp\left(-\frac{A_t - A}{A_c}\right), \text{ for } A_t \leq A, \quad (11)$$

Tsunamis: Stochastic Models of Occurrence and Generation Mechanisms,

Fig. 13 Comparison of Pareto cumulative distribution (blue line) with tapered and truncated Pareto cumulative distributions (red and green lines, respectively: dashed $A_c = 20$; solid $A_c = 200$). For all distributions, $\beta = 1$



where C is a constant (Kagan 1991, 2002). Figure 13 compares the pure, truncated, and tapered Pareto distributions using two different values for the corner/truncation size parameter. Parameter estimation methods for these distributions are described by Kagan (2002) and Meerschaert et al. (2012). Modified Pareto distributions have also been used to model the size distribution of landslides (Malamud et al. 2004; Stark and Hovius 2001).

Stochastic simulations based on a modified Pareto distribution can be used to generate synthetic tsunami catalogs. To generate a synthetic sample of an arbitrary length n using a Pareto distribution, the following expression is used

$$A = A_t R^{-1/\beta}, \quad (12)$$

where R is a random variable uniformly distributed on the interval $(0,1]$. For the tapered Pareto distribution, a synthetic sample is generated by selecting the minimum of the previous equation and

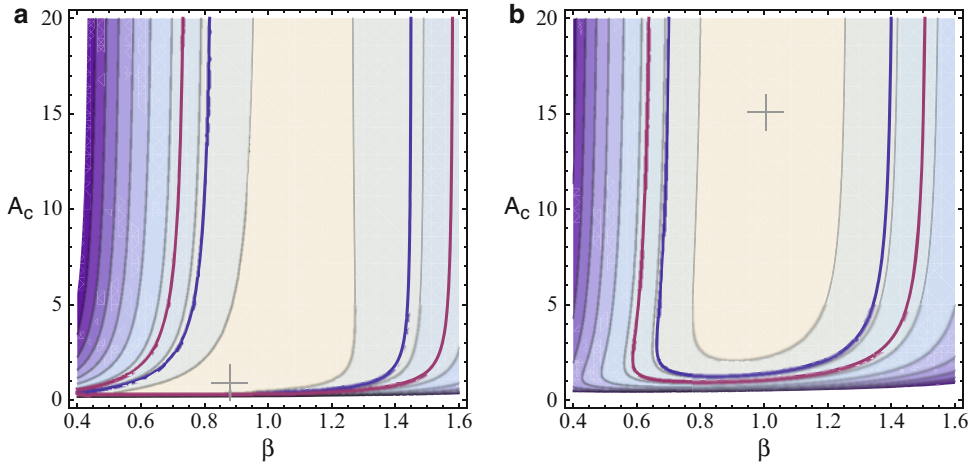
$$A = A_t - A_c \ln R \quad (13)$$

(Vere-Jones et al. 2001). Kagan (2002) describes simulation of random samples from other modified Pareto distributions. In terms of a marked point process, tsunami sizes generated from these stochastic simulations can provide the marks for a temporal ground process describing tsunami occurrence as discussed in the next section.

Tsunamis, as well as other natural hazards whose sizes are distributed according to a modified Pareto distribution, are undersampled by historical catalogs at their largest sizes. Geist and Parsons (2014) use stochastic simulations of tsunami catalogs as described above to show that the apparent size distribution derived from historical catalogs can appear either depleted of large events or by being characterized by one to a few large events. Parameter estimation tests indicate that whereas the Pareto exponent is well constrained by catalog data, estimates of the corner size parameter (A_c) associated with the tapered Pareto distribution are poorly determined, especially for the upper confidence limit (Geist and Parsons 2014). For example, parameter estimation results are shown in Fig. 14 for the Miyako tide gauge station in Japan that recorded the 2011 Tohoku tsunami. Whereas likelihood-based confidence intervals can be determined for the Pareto exponent (β), an upper confidence limit cannot be defined for the corner parameter (A_c). Obviously, these results have a significant effect on hazard assessments that rely on a worst-case scenario or specified by a low design probability.

Branching Process Models of Tsunami Source Occurrence

Similar to developing size distribution models for tsunamis described in the previous section, stochastic models for the occurrence of tsunami events in time are derived from those already



Tsunamis: Stochastic Models of Occurrence and Generation Mechanisms, Fig. 14 Two-parameter (β , A_c) likelihood contour maps for the Miyako tide gauge catalog (a) without and (b) with the 2011 events. Catalog lengths

are 51 and 52 events, respectively. Maximum likelihood estimates shown by plus sign. 95% and 99% confidence interval in parameter estimates shown by blue and magenta lines, respectively

developed and tested in the field of statistical seismology. There is evidence for temporal clustering of tsunami sources, both in terms of inter-event times and event counts (Geist and Parsons 2008, 2011). Because earthquakes are the primary generator of tsunamis, branching process models (Vere-Jones 2013) that have been developed to explain temporal clustering of earthquakes are immediately transferable. These models are classified as either branching-in-magnitude or branching-in-time (Kagan 2007, 2010).

Kagan (1973, 2010) developed the branching-in-magnitude model to simulate earthquake event counts. This model is based on a branching process including immigrants that is consistent with a Pareto distribution of earthquake sizes. The probability generating function for a specific cluster is that of a geometric distribution and a negative binomial distribution for a catalog of event clusters (Bartlett 1978; Kagan 2010; Kagan and Jackson 2000). The negative binomial distribution is an appropriate probability model if there is a large enough catalog of events in both space and time. Geist and Parsons (2011) adapt this branching-in-magnitude model for tsunami events. They indicate that in comparison to the negative binomial distribution, the Poisson null hypothesis can be rejected at the 95% confidence

level for tsunamis greater than 1 m in size from 1890 to 2010.

The most commonly invoked branching-in-time model for seismicity is the Epidemic Type Aftershock Sequence (ETAS) model developed by Ogata (1988). The ETAS model is based on the more general Hawkes self-exciting process, in which parent or spontaneous events are Poisson distributed in time with each successive Poisson-distributed generation following a Galton-Watson branching process (Daley and Vere-Jones 2003). The conditional intensity (Bebbington 2013) of the Hawkes self-exciting process over a history of events H_t is given by

$$\lambda(t|H_t) = \mu + \sum_{i:t_i < t} g(t - t_i) \quad (14)$$

where $\mu > 0$ is the background rate and $g(t - t_i) \geq 0$ is the triggering function. For the ETAS model, the triggering function is specified by the Omori-Utsu temporal distribution of aftershocks, such that

$$\lambda(t|H_t) = \mu + \sum_{i:t_i < t} \frac{K_i}{(t - t_i + c)^p}, \quad (15)$$

where c represents a limit where triggered events can be detected and p is an exponent that controls

the temporal decay of aftershocks. Parsons (2002) indicates that the Omori-Utsu distribution holds more generally for large-magnitude triggered events beyond just aftershocks on the same fault. K_i is the productivity of triggered events and is dependent on the magnitude of earthquake i (Ogata 1988):

$$K_i = K_0 e^{\alpha(M_i - M_t)}, \quad (16)$$

where K_0 is a normalizing constant, M_t is a lower magnitude threshold, and α is a coefficient for the magnitude effect on triggering.

The primary result of estimating ETAS parameters from the tsunami event catalog is that the magnitude scaling of triggered events (parameter α) is essentially zero, in contrast to non-zero magnitude scaling for earthquake catalogs of identical magnitude thresholds (Geist 2014). If tsunamigenic conditions (i.e., submarine location, shallow focal depth, dip-slip mechanism, in addition to the magnitude threshold) are applied to earthquake catalogs, ETAS parameter estimates are similar to those derived from tsunami event catalogs. It appears that the dip-slip mechanism is the cause of zero magnitude scaling in the triggering function (Geist 2014).

Stochastic Parameter Estimation of Paleotsunami Data

For paleotsunami data (both submarine sources and tsunami deposits onshore), several methods have been developed to estimate the occurrence probability of tsunamis, taking into account age-dating uncertainty of individual horizons that represent tsunami or source events. In addition, one has to account for the open time intervals before the first and after the last observed event. The probability estimates derived from paleotsunami data apply to all tsunami sizes associated with the events exposed at a particular site.

In many cases, particularly for submarine landslide tsunami sources, the dates of individual events are unknown, although a potentially datable geological horizon exists below a sequence of events imaged by seismic reflection profiles. Assuming an exponential distribution of inter-

event times associated with a stationary Poisson process, the rate parameter (λ) can be treated as a random variable, and its uncertainty can be determined. Bayesian methods are particularly well suited to estimating the empirical probability and associated uncertainty from small sample sizes. In the empirical Bayes approach, a conjugate prior distribution is used, i.e., the resulting posterior distribution is in the same distribution family (e.g., the exponential distribution family) as the prior distribution. The Poisson-Gamma conjugate prior model is used by Geist and Parsons (2010) to estimate λ for submarine landslides, and the beta-binomial conjugate prior model is used by Savage (1994) for determining earthquake probability from paleoseismic horizons.

For the case where individual events are dated, non-Poisson distributions can be evaluated for clustering or quasiperiodic behavior. Techniques used to test different probability models include maximum likelihood, Monte Carlo, and type II maximum likelihood, as summarized by Geist et al. (2013). Of note, both the Monte Carlo (Parsons 2008) and the Bayesian type II maximum likelihood technique (Ogata 1999) account for both the age-dating uncertainty and the open time intervals. Unlike traditional Bayesian inference techniques in which the prior and posterior distribution are a function of the parameter space (θ), the type II maximum likelihood method is formulated such that the prior and posterior distributions are the probability densities of each event age distribution ($\psi_i(t_i)$; $i = 1, \dots, n$) over the time interval $[S, T]$. Thus, both the event ages and models parameters are uncertain. Bayes' theorem is then written as

$$\varphi(t_1, \dots, t_n | \theta) = \frac{L_{[S,T]}(\theta) \prod_{i=1}^n \psi_i(t_i)}{\Lambda_{[S,T]}(\theta)} \quad (17)$$

where $\varphi(t_1, \dots, t_n | \theta)$ is the posterior distribution, L is the likelihood function, and $\Lambda_{[S,T]}(\theta)$ is termed the integrated or marginal likelihood (Aitkin 2010; Ogata 1999). $\Lambda_{[S,T]}(\theta)$ integrates the likelihood function with the joint distribution of the priors:

$$\Lambda_{[S,T]}(\theta) = \int_S \dots \int_S^T L_{[S,T]}(\theta; t_1, \dots, t_n) \prod_{i=1}^n \psi_i(t_i) dt_1, \dots, dt_n \quad (18)$$

Ogata (1999) provides details on the formulation of $\Lambda_{[S,T]}(\theta)$. Akaike's Bayesian information criterion (ABIC) (Akaike 1980) can then be used to select the optimal model from this method.

Probabilistic Analysis of Tsunami Hazards

PTHA Framework

The stochastic methods described above can be applied to probabilistic tsunami hazard analysis (PTHA). Adapted from probabilistic seismic hazard analysis, PTHA aggregates tsunami run-up or amplitude at a site from many different source locations and source parameter combinations as described by Grezio et al. (2017), who also summarize recent developments in PTHA methodologies. It is assumed that tsunamis occur as a stationary Poisson process. The framework PTHA equation for the rate parameter indicated by Geist et al. (2009) is given by

$$\lambda(R > R_0) = \sum_{\text{type}=i} \sum_{\text{zone}=j} v_{ij} \int P(R > R_0 | \psi_{ij}) f_{\psi}(\psi_{ij}) d\psi \quad (19)$$

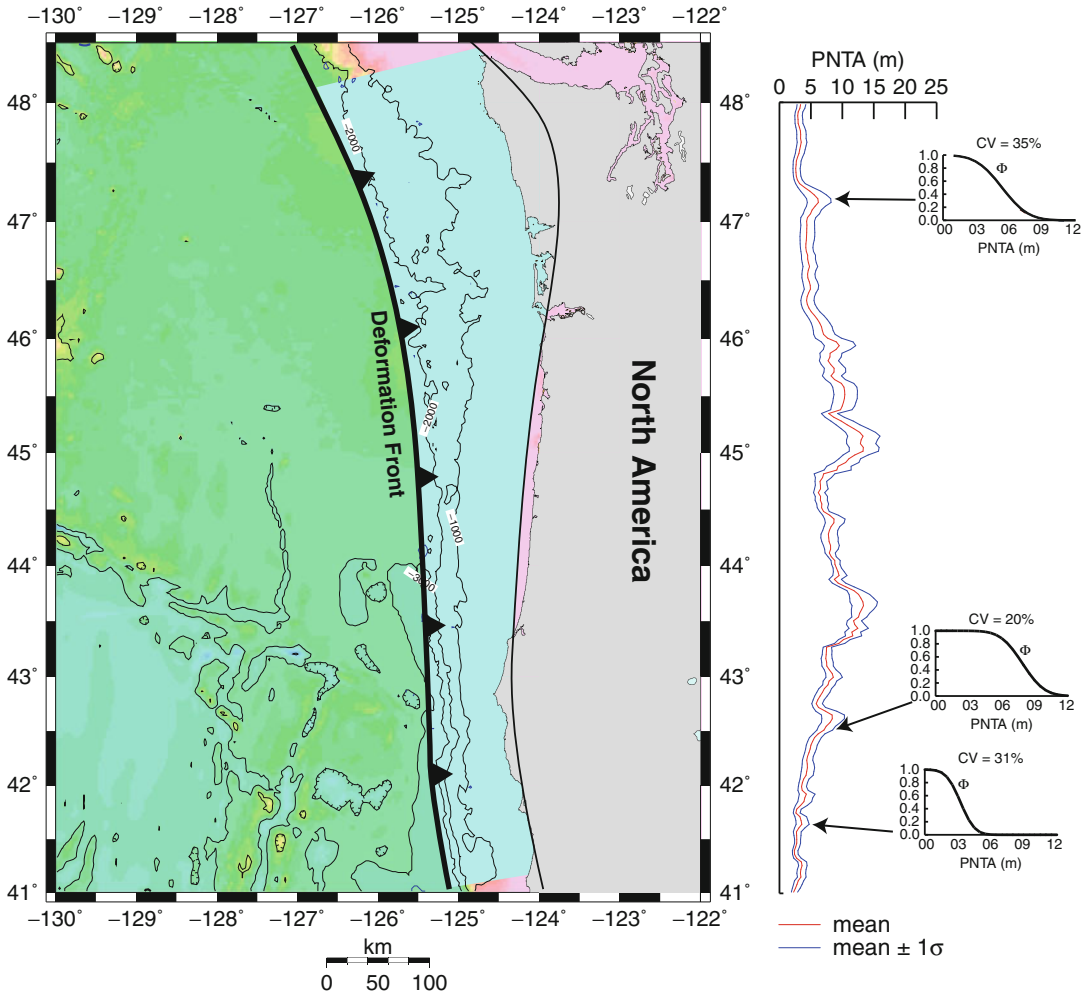
where $P(R > R_0 | \psi_{ij})$ is the probability that tsunami amplitude or run-up exceeds a threshold value (R_0), given a set of source parameters (ψ) for each source type (i) and source zone (j). $P(R > R_0 | \psi_{ij})$ is computed using numerical propagation modeling that explicitly includes distance attenuation and propagation path effects. Typically, the distribution $f_{\psi}(\psi_{ij})$ is set equal to the size distribution of tsunami source (e.g., seismic moment distribution for earthquakes, volume or area distribution for landslides, etc.), such that $P(R > R_0 | \psi_{ij})$ includes uncertainty in scaling relationships between individual source parameters and the primary size variable. For example, various scaling relationships between earthquake magnitude and rupture area and mean slip have

recently been proposed for subduction zone earthquakes (Allen and Hayes 2017; Blaser et al. 2010; Murotani et al. 2013; Strasser et al. 2010). Goda et al. (2016) specifically derive source parameters in reference to stochastic slip models used in tsunami generation. Other sources of uncertainty incorporated into PTHA are described by Grezio et al. (2017).

Modified Pareto distributions have typically been used for $f_{\psi}(\psi_{ij})$ in the past (Parsons and Geist 2009). Because numerical methods are used to compute the aggregate rate parameter, discrete values of source sizes are needed. Sampling of a Pareto distribution using the methods described above provides the seismic moment for which tsunami generation and propagation models are based. Rather than using a modified Pareto distribution for $f_{\psi}(\psi_{ij})$, several PTHA studies have incorporated characteristic distributions where only the largest earthquake (not the Pareto-distributed smaller earthquakes, such as aftershocks of the characteristic event) is used (Annaka et al. 2007). The danger in this approach is that the tsunami hazard at rates higher than that for the characteristic earthquake would be underestimated. In addition, the characteristic assumption does not account for a possibility of tsunamigenic earthquakes occurring with magnitudes larger than that of the characteristic earthquake. A logic tree is often used in probabilistic hazard analysis studies to incorporate different scientific interpretations of source characterization (Annaka et al. 2007; Burbridge et al. 2008; Power et al. 2013). In addition, although Selva et al. (2016) provide an ensemble modeling approach especially suited to PTHA calculations as a useful alternative to logic trees.

Implementing Stochastic Source Models into PTHA

A primary source of aleatory uncertainty that can be included in PTHA is related to slip distribution. Stochastic source models are important tools used to estimate this uncertainty. As an example in Fig. 15, the probability $P(R > R_0 | \psi_{ij})$ is shown for a Cascadia earthquake rupture ($M_w = 9$) in which 100 realizations of the Herrero and Bernard (1994) static stochastic slip model are used (Geist



Tsunamis: Stochastic Models of Occurrence and Generation Mechanisms, Fig. 15 (Left) Map of Cascadia subduction zone showing $M_w = 9$ Cascadia earthquake. Light blue: rupture area. Bathymetric contour interval: 1000 m. (Right) Mean and $\pm 1\sigma$ values of peak nearshore

tsunami amplitude (PNTA) curves and cumulative probability curves (Φ) resulting from 100 stochastic slip distributions. CV = coefficient of variation. See Geist and Parsons (2006) for details of computation

and Parsons 2006). Because the $P(R > R_0 | \psi_{ij})$ curves include propagation effects (but no other source parameter uncertainty), they vary considerably along the Pacific Northwest coastline. Results for Pacific Northwest tsunami hazards from Cascadia earthquakes have been refined significantly by LeVeque et al. (2016) who account for the non-planar geometry of the subduction zones. In addition, Lorito et al. (2015) and Sepúlveda et al. (2017) develop strategies for significantly reducing the computational cost of including slip uncertainty into PTHA calculations.

Regional PTHA studies incorporating stochastic slip include those for Indonesia (Griffin et al. 2017; Horspool et al. 2014), New Zealand (Mueller et al. 2015), Japan (Goda et al. 2015a; Goda and Song 2016), and the USA (González et al. 2009).

Role of Dynamic Rupture Models in PTHA Calculations

It is in dealing with the uncertainty in tsunami amplitude that dynamic modeling may help guide our efforts. Returning to our dynamic modeling example from the Stochastic Models

of Tsunami Generation section, we run ten different randomization realizations and then use the resulting (time-dependent) surface deformations as sources for ten different calculations of near-shore tsunami amplitude along the Japanese coast, using the local bathymetry. The results of these calculations are shown in Fig. 16a.

These results show quite clearly the variability in tsunami amplitude that results from both the spatial variations in fault slip and the presence of the splay fault. Much of the variability along strike for a given realization can be attributed to propagation effects and the local seafloor and shoreline topography. However, in Fig. 16a we see a large variation among different randomized stress events, especially in the region of the shoreline closest to the fault (between 550 and 625 km along strike). This effect can be seen in both the peak nearshore tsunami amplitude and the coefficient of variation for all the models. Note that there is a large central peak in the tsunami amplitude for only some of the models; these are the ones that significantly activate the splay fault. Most models do not activate the splay significantly and do not produce this peak. However, even when rupture proceeds onto the splay fault, there is significant variation in the amount and distribution of splay fault slip. Thus, we immediately see that the presence of complex fault geometry can greatly increase variability in tsunami generation – something that is typically not accounted for in standard kinematic source models.

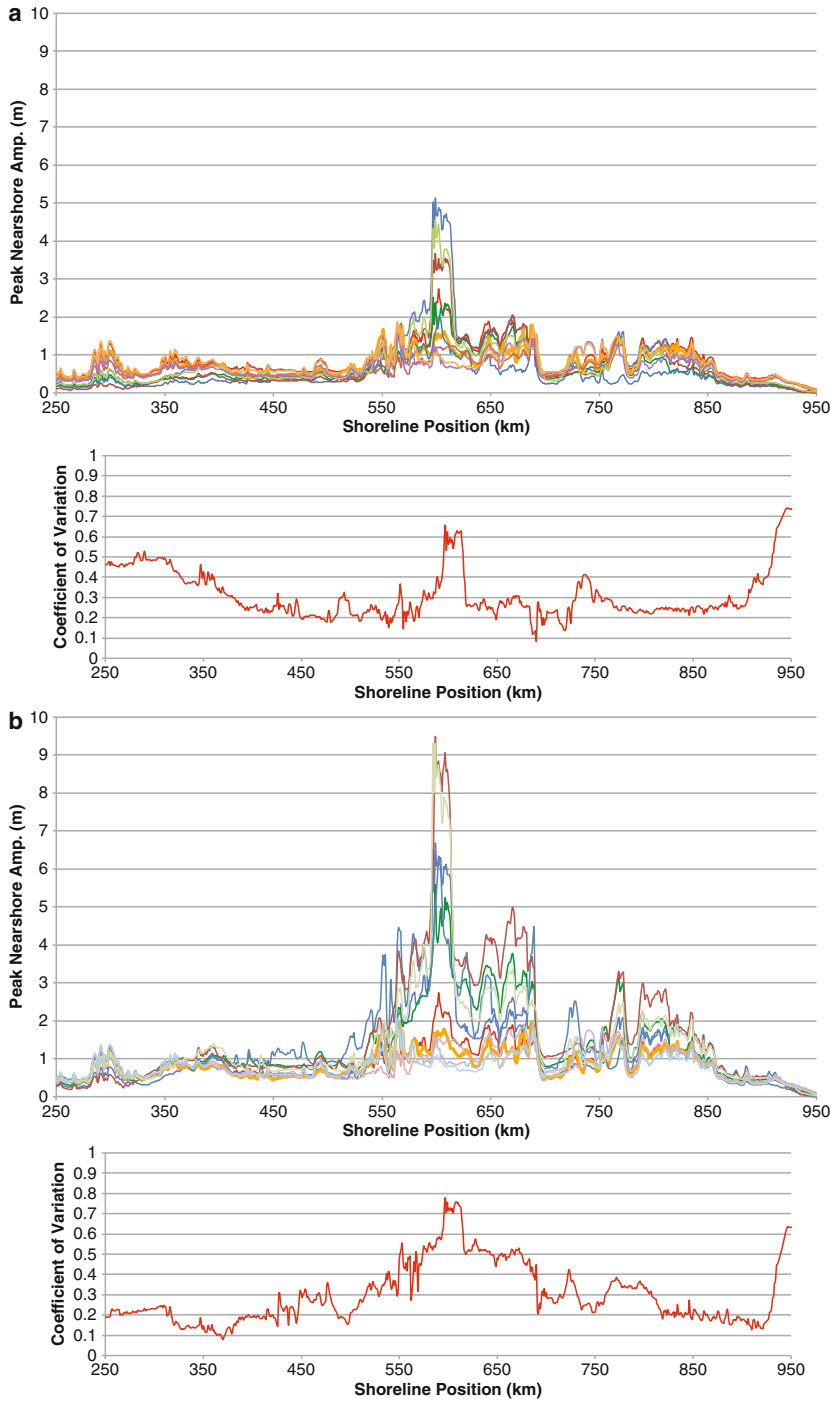
As noted above, one key difference between dynamic models and kinematic models is that the overall magnitude of the event is a calculated result of the modeling process. However, as noted above in the Dynamic Rupture Models of Earthquakes section, recent efforts have been made to generate stochastic prestress that encompasses an approximate earthquake magnitude beforehand on rectangular and planar strike-slip faults (Andrews and Ma 2016). The earthquake sources in Fig. 16a do not have the same seismic moment; they vary from 6.4×10^{20} to 2.0×10^{21} Nm. Thus, some of the variability between the generated tsunamis could be related to this variability in source magnitude. To compensate for this effect, we run a second suite of

dynamic rupture models using the same stress fields as before, but scaled upwards to produce as equivalent final seismic moments as possible as a calculation result. These results are shown in Fig. 16b, in which the source moment varies only from 1.88×10^{21} to 2.11×10^{21} Nm. Significantly, the scaled models produce slightly lower variability in the regions farthest from the fault (where the tsunami amplitude is more likely to be dominated purely by the earthquake size), but the variability actually increased in the near-source region. This counterintuitive result is due to the fact that on the whole, earthquakes that activated the splay fault tended to have lower seismic moments than those that did not, while still producing the highest near-source tsunami amplitude. Thus, when these splay-rupturing earthquake stress fields were scaled up to match the other earthquakes, their tsunami amplitudes scaled up accordingly.

Dynamic models such as those shown above are computationally expensive and may not be feasible directly for PTHA use, in which hundreds of models or more may be necessary. Nonetheless, they may help to narrow down the regions of parameter space that are physically plausible and lead to better constructed kinematic models (such as guiding the partitioning of slip between splays and megathrust faults), as well as indicating the likelihood of rupture propagating through barriers (such as low seismic coupling zones) on faults. Additionally, using time-dependent surface generation from dynamic rupture models as input into tsunami models can help shed light on near-source variability. In the future, further applications of dynamic modeling may help more accurately encompass tsunami generation scenarios among complex fault systems based on physical rather than historical arguments.

Future Directions

Stochastic models have only recently been introduced to describe the generation and evolution of tsunamis and to analyze patterns of occurrence for tsunami events. The most common application of stochastic methods for tsunamis is in the area of specifying heterogeneous slip for tsunami



Tsunamis: Stochastic Models of Occurrence and Generation Mechanisms, Fig. 16 Peak nearshore tsunami amplitude and coefficient of variation resulting from ten stochastic stress realizations. (a) Raw results, for which there is a variation in earthquake seismic moment. (b) Results for dynamic models with stresses scaled to produce

equivalent seismic moments. Two sample slip distributions resulting from the dynamic model are shown in Fig. 4; the right model in Fig. 4, which has significant slip on the splay fault, produces a high central peak in Fig. 13; the left model does not

generation. Although most of these applications have been focused toward tsunami hazard assessment, recent studies have shown promise in applying stochastic slip models to tsunami forecasts as well, in terms of quantification of uncertainty (Babeyko et al. 2010; Blaser et al. 2011; Cienfuegos et al. 2018). In the future, stochastic models of seismogenic tsunami generation that specify static slip on the fault may in many cases be replaced by dynamic rupture propagation models in the presence of stochastic stress. Dynamic rupture models will continue to be developed in the future and will be important in providing an understanding of rupture paths on complex fault systems, such as those observed in subduction zones. In the short term, dynamic models may become useful in influencing the construction of more computationally efficient kinematic models by allowing the evaluation of correlations between source parameters such as slip, rupture velocity, and rupture path (cf., Guatteri et al. 2004).

Analysis of tsunami time series and tsunami catalogs has also been examined as stochastic processes. During propagation, the tsunami wavefield becomes extremely complex owing to reflections, scattering, refractive focusing and defocusing, etc. Methods adapted from both scattering theory in seismology (Saito and Furumura 2009) and wave statistics for a random sea in oceanography (Geist 2009) have recently been successful in explaining this complexity. Similarly, methods developed in statistical seismology are easily adapted to analyze tsunami event catalogs, in terms of the distribution of sizes and the occurrence of tsunami events in time. Advances in oceanographic wave theory and statistical seismology will be critical in defining uncertainty for both deterministic and probabilistic tsunami assessments in the future.

Acknowledgment The authors gratefully acknowledge careful readings of the first edition of this paper and constructive comments by Martin Mai, Mark Bebbington, Kenji Satake, Tom Parsons, and the Encyclopedia Section Editor William H. K. Lee. Data used in this study include DART bottom-pressure recorder data from the event archive at NOAA's National Data Buoy Center and digital tide gauge records from NOAA's National Tsunami

Warning Center. Analysis of tsunami records was performed in part using the Wave Analysis for Fatigue and Oceanography (WAFO) package developed at Lund University, Sweden.

Bibliography

- Aitkin M (2010) Statistical inference: an integrated Bayesian/likelihood approach, vol 116. CRC Press, Boca Raton, p 236
- Akaike H (1980) Likelihood and the Bayes procedure. *Trab Estad Investig Oper* 31:143–166
- Aki K, Chouet B (1975) Origin of coda waves: source, attenuation, and scattering effects. *J Geophys Res* 80:3322–3342
- Allen TI, Hayes GP (2017) Alternative rupture-scaling relationships for subduction Interface and other offshore environments. *Bull Seismol Soc Am* 107:1240–1253
- Andrews DJ (1976) Rupture velocity of plane strain shear cracks. *J Geophys Res* 81:5679–5687
- Andrews DJ (1980) A stochastic fault model 1. Static case. *J Geophys Res* 85:3867–3877
- Andrews DJ, Barall M (2011) Specifying initial stress for dynamic heterogeneous earthquake source models. *Bull Seismol Soc Am* 101:2408–2417
- Andrews DJ, Ma S (2016) Validating a dynamic earthquake model to produce realistic ground motion. *Bull Seismol Soc Am* 106:665–672
- Annaka T, Satake K, Sakakiyama T, Yanagisawa K, Shuto N (2007) Logic-tree approach for probabilistic tsunami hazard analysis and its applications to the Japanese coasts. *Pure Appl Geophys* 164:577–592
- Aochi H, Fukuyama E, Matsu'ura M (2000) Selectivity of spontaneous rupture propagation on a branched fault. *Geophys Res Lett* 27:3635–3638
- Babeyko A, Hoechner A, Sobolev S (2010) Source modeling and inversion with near real-time GPS: a GITEWS perspective for Indonesia. *Nat Hazards Earth Syst Sci* 10:1617
- Bartlett MS (1978) An introduction to stochastic processes, 3rd edn. Cambridge University Press, Cambridge, p 388
- Bebbington M (2013) Volcanic eruptions: stochastic models of occurrence patterns. In: Meyers R (ed) *Encyclopedia of complexity and systems science*. Springer, Berlin/Heidelberg
- Beichelt FE, Fatti LP (2002) Stochastic processes and their applications. Taylor & Francis, London
- Blaser L, Krüger F, Ohrnberger M, Scherbaum F (2010) Scaling relations of earthquake source parameter estimates with special focus on subduction environment. *Bull Seismol Soc Am* 100:2914–2926
- Blaser L, Ohrnberger M, Riggelsen C, Babeyko A, Scherbaum F (2011) Bayesian networks for tsunami early warning. *Geophys J Int* 185:1431–1443
- Bormann P, Aki K, Lee WHK, Schweitzer J (2013) Glossary. In: Bormann P (ed) *New manual of seismological*

- observatory practice 2 (NMSOP2). Deutsches GeoForschungsZentrum GFZ, Potsdam, pp 1–200. https://doi.org/10.2312/GFZ.NMSOP-2_Glossary
- Burbridge D, Cummins PR, Mleczko R, Thio HK (2008) A probabilistic tsunami hazard assessment for western Australia. *Pure Appl Geophys* 165:2059–2088
- Burroughs SM, Tebbens SF (2005) Power law scaling and probabilistic forecasting of tsunami runup heights. *Pure Appl Geophys* 162:331–342
- Candela T, Renard F, Klinger Y, Mair K, Schmittbuhl J, Brodsky EE (2012) Roughness of fault surfaces over nine decades of length scales. *J Geophys Res* 117. <https://doi.org/10.1002/2011JB009041>
- Carrier GF (1970) Stochastically driven dynamical systems. *J Fluid Mech* 44:249–264
- Carrier GF (1995) On-shelf tsunami generation and coastal propagation. In: Tsuchiya Y, Shuto N (eds) *Tsunami: Progress in prediction, disaster prevention and warning*, vol 4. Kluwer, Dordrecht, pp 1–20
- Cienfuegos R, Catalán PA, Urrutia A, Benavente R, Aránguiz R, González G (2018) What can we do to forecast tsunami hazards in the near field given large epistemic uncertainty in rapid seismic source inversions? *Geophys Res Lett* 45:4944–4955
- Corral A (2009) Point-occurrence self-similarity in crackling-noise systems and in other complex systems. *J Stat Mech Theor Exp* 2009. <https://doi.org/10.1088/1742-5468/2009/01/P01022>
- Daley DJ, Vere-Jones D (2003) *An introduction to the theory of point processes*, vol 1, 2nd edn. Springer, New York
- Das S (1981) Three-dimensional spontaneous rupture propagation and implications for the earthquake source mechanism. *Geophys J R Astron Soc* 67:375–393
- Davies G, Horspool N, Miller V (2015) Tsunami inundation from heterogeneous earthquake slip distributions: evaluation of synthetic source models. *J Geophys Res Solid Earth* 120:6431–6451
- Day SM (1982) Three-dimensional simulation of spontaneous rupture: the effect of nonuniform prestress. *Bull Seismol Soc Am* 72:1881–1902
- DeDontney N, Rice JR, Dmowska R (2012) Finite element modeling of branched ruptures including off-fault plasticity. *Bull Seismol Soc Am* 102:541–562
- Duan B, Oglesby DD (2007) Nonuniform prestress from prior earthquakes and the effect on dynamics of branched fault systems. *J Geophys Res* 112:B05308. <https://doi.org/10.1029/2006JB004443>
- Fukutani Y, Suppasri A, Imamura F (2015) Stochastic analysis and uncertainty assessment of tsunami wave height using a random source parameter model that targets a Tohoku-type earthquake fault. *Stoch Env Res Risk A* 29:1763–1779
- Geist EL (2002) Complex earthquake rupture and local tsunamis. *J Geophys Res* 107. <https://doi.org/10.1029/2000JB000139>
- Geist EL (2009) Phenomenology of tsunamis: statistical properties from generation to runup. *Adv Geophys* 51:107–169
- Geist EL (2012a) Near-field tsunami edge waves and complex earthquake rupture. *Pure Appl Geophys*. <https://doi.org/10.1007/s00024-012-0491-7>
- Geist EL (2012b) Phenomenology of tsunamis II: scaling, event statistics, and inter-event triggering. *Adv Geophys* 53:35–92
- Geist EL (2014) Explanation of temporal clustering of tsunami sources using the epidemic-type aftershock sequence model. *Bull Seismol Soc Am* 104:2091–2103
- Geist EL (2016) Non-linear resonant coupling of tsunami edge waves using stochastic earthquake source models. *Geophys J Int* 204:878–891
- Geist EL (2018) Effect of dynamical phase on the resonant interaction among tsunami edge wave modes. *Pure Appl Geophys* 175:1341–1354
- Geist EL, Oglesby DD (2014) Earthquake mechanism and seafloor deformation for tsunami generation. In: Beer M, Patelli E, Kougiumtzooglou IA, Au IS-K (eds) *Encyclopedia of earthquake engineering*. Springer, Berlin, p 17
- Geist EL, Parsons T (2006) Probabilistic analysis of tsunami hazards. *Nat Hazards* 37:277–314
- Geist EL, Parsons T (2008) Distribution of tsunami inter-event times. *Geophys Res Lett* 35:L02612. <https://doi.org/10.1029/2007GL032690>
- Geist EL, Parsons T (2010) Estimating the empirical probability of submarine landslide occurrence. In: Mosher DC, Shipp C, Moscardelli L, Chaytor J, Baxter C, Lee HJ, Urgeles R (eds) *Submarine mass movements and their consequences IV*. Springer, Heidelberg, pp 377–386
- Geist EL, Parsons T (2011) Assessing historical rate changes in global tsunami occurrence. *Geophys J Int* 187:497–509
- Geist EL, Parsons T (2014) Undersampling power-law size distributions: effect on the assessment of extreme natural hazards. *Nat Hazards* 72:565–595
- Geist EL, Parsons T, ten Brink US, Lee HJ (2009) Tsunami Probability. In: Bernard EN, Robinson AR (eds) *The sea*, vol 15. Harvard University Press, Cambridge, pp 93–135
- Geist EL, Chaytor JD, Parsons T, ten Brink U (2013) Estimation of submarine mass failure probability from a sequence of deposits with age dates. *Geosphere* 9:287–298
- Goda K, Song J (2016) Uncertainty modeling and visualization for tsunami hazard and risk mapping: a case study for the 2011 Tohoku earthquake. *Stoch Env Res Risk A* 30:2271–2285
- Goda K, Mai PM, Yasuda T, Mori N (2014) Sensitivity of tsunami wave profiles and inundation simulations to earthquake slip and fault geometry for the 2011 Tohoku earthquake. *Earth Planets Space* 66:105
- Goda K, Li S, Mori N, Yasuda T (2015a) Probabilistic tsunami damage assessment considering stochastic source models: application to the 2011 Tohoku earthquake. *Coast Eng J* 57:1550015
- Goda K, Yasuda T, Mori N, Mai PM (2015b) Variability of tsunami inundation footprints considering stochastic

- scenarios based on a single rupture model: application to the 2011 Tohoku earthquake. *J Geophys Res Oceans* 120:4552–4575
- Goda K, Yasuda T, Mori N, Maruyama T (2016) New scaling relationships of earthquake source parameters for stochastic tsunami simulation. *Coast Eng J* 58:1650010–1–1650010–40
- González FI, Kulikov YA (1993) Tsunami dispersion observed in the deep ocean. In: Tinti S (ed) *Tsunamis in the world*. Kluwer Academic Publishers, Dordrecht, pp 7–16
- González FI, Geist EL, Jaffe BE, Kânoglu U, Mofjeld HO, Synolakis CE et al (2009) Probabilistic tsunami hazard assessment at seaside, Oregon for near- and far-field seismic sources. *J Geophys Res* 114. <https://doi.org/10.1029/2008JC005132>
- Greenough J, Main IG (2008) A Poisson model for earthquake frequency uncertainties in seismic hazard analysis. *Geophys Res Lett* 35. <https://doi.org/10.1029/2008GL035353>
- Grezio A, Babeyko AY, Baptista AM, Behrens J, Costa A, Davies G et al (2017) Probabilistic tsunami Hazard analysis (PTHA): multiple sources and global applications. *Rev Geophys* 55:1158–1198
- Griffin JD, Pranantyo IR, Kongko W, Haunan A, Robiana R, Miller V et al (2017) Assessing tsunami hazard using heterogeneous slip models in the Mentawai Islands, Indonesia. *Geol Soc Lond, Spec Publ* 441:47–70
- Gutteri M, Mai PM, Beroza GC (2004) A pseudo-dynamic approximation to dynamic rupture models for strong ground motion prediction. *Bull Seismol Soc Am* 94:2051–2063
- Gutenberg B, Richter CF (1944) Frequency of earthquakes in California. *Bull Seismol Soc Am* 34: 185–188
- Harris RA, Archuleta RJ, Day SM (1991) Fault steps and the dynamic rupture process – 2-D numerical simulations of a spontaneously propagating shear fracture. *Geophys Res Lett* 18:893–896
- Hergarten S (2003) Landslides, sandpiles, and self-organized criticality. *Nat Hazards Earth Syst Sci* 3:505–514
- Hergarten S, Neugebauer HJ (1998) Self-organized criticality in a landslide model. *Geophys Res Lett* 25:801–804
- Herrero A, Bernard P (1994) A kinematic self-similar rupture process for earthquakes. *Bull Seismol Soc Am* 84:1216–1228
- Hisada Y (2000) A theoretical omega-square model considering the spatial variation in slip and rupture velocity. *Bull Seismol Soc Am* 90:387–400
- Hisada Y (2001) A theoretical omega-square model considering the spatial variation in slip and rupture velocity. Part 2: case for a two-dimensional source model. *Bull Seismol Soc Am* 91:651–666
- Hogben N (1990) Long term wave statistics. In: Le Méhauté B, Hanes DM (eds) *Ocean engineering science*, vol 9. Wiley, New York, pp 293–333
- Horspool N, Pranantyo I, Griffin J, Latief H, Natawidjaja DH, Kongko W et al (2014) A probabilistic tsunami hazard assessment for Indonesia. *Nat Hazards Earth Syst Sci* 14:3105–3122
- Huang NE, Tung CC, Long SR (1990) The probability structure of the ocean surface. In: Le Méhauté B, Hanes DM (eds) *Ocean engineering science*, vol 9. Wiley, New York, pp 335–366
- Ishimoto M, Iida K (1939) Observations of earthquakes registered with the microseismograph constructed recently. *Bull Earthquake Res Inst* 17:443–478
- Jha AK, Winterstein SR (2000) Nonlinear random ocean waves: prediction and comparison with data. In: *Proceedings of the ETCE/OMAE2000 joint conference: energy for the new Millennium*. ASME, New Orleans, pp 1–12
- Kagan YY (1973) Statistical methods in the study of seismic processes. *Bull Int Stat Inst* 45:437–453
- Kagan YY (1991) Seismic moment distribution. *Geophys J Int* 106:123–134
- Kagan YY (2002) Seismic moment distribution revisited: I. statistical results. *Geophys J Int* 148: 520–541
- Kagan YY (2007) Why does theoretical physics fail to explain and predict earthquake occurrence? In: Bhattacharyya P, Chakrabarti BK (eds) *Modelling critical and catastrophic phenomena in geoscience*. Springer, Berlin, pp 303–359
- Kagan YY (2010) Statistical distributions of earthquake numbers: consequence of branching process. *Geophys J Int* 180:1313–1328
- Kagan YY, Jackson DD (2000) Probabilistic forecasting of earthquakes. *Geophys J Int* 143:438–453
- Kame N, Rice JR, Dmowska R (2003) Effects of pre-stress state and rupture velocity on dynamic fault branching. *J Geophys Res* 108:2265. <https://doi.org/10.1029/2002JB002189>
- Kanamori H (1977) The energy release in great earthquakes. *J Geophys Res* 82:2981–2987
- Kirby JT, Putrevu U, Özkan-Haller HT (1998) Evolution equations for edge waves and shear waves on longshore uniform beaches. In: *Proceedings of 26th international conference on coastal engineering*. ASCE, Copenhagen, pp 203–216
- Koshimura S, Hayashi Y, Munemoto K, Imamura F (2008) Effect of the emperor seamounts on trans-oceanic propagation of the 2006 Kuril Island earthquake tsunami. *Geophys Res Lett* 35. <https://doi.org/10.1029/2007GL032129>
- Kowalik Z, Horrillo J, Knight W, Logan T (2008) Kuril Islands tsunami of November 2006: 1. Impact at Crescent City by distant scattering. *J Geophys Res* 113. <https://doi.org/10.1029/2007JC004402>
- Lavallée D (2008) On the random nature of earthquake sources and ground motions: a unified theory. *Adv Geophys* 50:427–461
- Lavallée D, Liu P, Archuleta RJ (2006) Stochastic model of heterogeneity in earthquake slip spatial distributions. *Geophys J Int* 165:622–640

- Lavallée D, Miyake H, Koketsu K (2011) Stochastic model of a subduction-zone earthquake: sources and ground motions for the 2003 Tokachi-oki, Japan, earthquake. *Bull Seismol Soc Am* 101:1807–1821
- Lay T, Kanamori H, Ammon CJ, Hutto AR, Furlong K, Rivera L (2009) The 2006–2007 Kuril Islands great earthquake sequence. *J Geophys Res* 114. <https://doi.org/10.1029/2008JB006280>
- LeVeque RJ, Waagan K, González FI, Rim D, Lin G (2016) Generating random earthquake events for probabilistic tsunami Hazard assessment. *Pure Appl Geophys* 173:3671–3692
- Liu-Zeng J, Heaton TH, DiCaprio C (2005) The effect of slip variability on earthquake slip-length scaling. *Geophys J Int* 162:841–849
- Longuet-Higgins MS (1952) On the statistical distribution of the heights of sea waves. *J Mar Res* 11:245–266
- Lorito S, Selva J, Basili R, Romano F, Tiberti MM, Piatanesi A (2015) Probabilistic hazard for seismically induced tsunamis: accuracy and feasibility of inundation maps. *Geophys J Int* 200:574–588
- Løvholt F, Pedersen G, Bazin S, Kühn D, Bredesen RE, Harbitz C (2012) Stochastic analysis of tsunami runup due to heterogeneous coseismic slip and dispersion. *J Geophys Res* 117. <https://doi.org/10.1029/2011JC007616>
- Ma S, Beroza GC (2008) Rupture dynamics on a bimaterial interface for dipping faults. *Bull Seismol Soc Am* 98:1642–1658
- Mai PM, Beroza GC (2002) A spatial random field model to characterize complexity in earthquake slip. *J Geophys Res* 107. <https://doi.org/10.1029/2001JB000588>
- Main I (2000) Apparent breaks in scaling in the earthquake cumulative frequency-magnitude distribution: fact or artifact? *Bull Seismol Soc Am* 90:86–97
- Malamud BD, Turcotte DL, Guzzetti F, Reichenbach P (2004) Landslide inventories and their statistical properties. *Earth Surf Process Landf* 29:687–711
- Massel SR (1996) Ocean surface waves: their physics and prediction, vol 11. World Scientific, Singapore
- McCloskey J, Antonioli A, Piatanesi A, Sieh K, Steacy S, Nalbant S et al (2008) Tsunami threat in the Indian Ocean from a future megathrust earthquake west of Sumatra. *Earth Planet Sci Lett* 265:61–81
- McKane AJ (2016) Stochastic processes. In: Meyers RA (ed) *Encyclopedia of complexity and systems science*. Springer, Berlin/Heidelberg, pp 1–22
- Meerschaert MM, Roy P, Shao Q (2012) Parameter estimation for exponentially tempered power law distributions. *Commun Stat Theor Models* 41:1839–1856
- Mei CC, Stiassnie M, Yue DK-P (2005) Theory and applications of ocean surface waves. Part 2: nonlinear aspects. World Scientific, Singapore
- Mikumo T, Miyatake T (1993) Dynamic rupture processes on a dipping fault, and estimates of stress drop and strength excess from the results of waveform inversion. *Geophys J Int* 112:481–496
- Mofjeld HO, González FI, Bernard EN, Newman JC (2000) Forecasting the heights of later waves in Pacific-wide tsunamis. *Nat Hazards* 22:71–89
- Mofjeld HO, Titov VV, González FI, Newman JC (2001) Tsunami scattering provinces in the Pacific Ocean. *Geophys Res Lett* 28:335–337
- Mueller C, Power W, Fraser S, Wang X (2015) Effects of rupture complexity on local tsunami inundation: implications for probabilistic tsunami hazard assessment by example. *J Geophys Res Solid Earth* 120:488–502
- Muraleedharan G, Rao AD, Kurup PG, Nair NU, Sinha M (2007) Modified Weibull distribution for maximum and significant wave height simulation and prediction. *Coast Eng* 54:630–638
- Murotani S, Satake K, Fujii Y (2013) Scaling relations of seismic moment, rupture area, average slip, and asperity size for M₀–9 subduction-zone earthquakes. *Geophys Res Lett* 40:5070–5074
- Ochi MK (2005) *Ocean waves: the stochastic approach*. Cambridge University Press, Cambridge, p 332
- Ogata Y (1988) Statistical models for earthquake occurrences and residual analysis for point processes. *J Am Stat Assoc* 83:9–27
- Ogata Y (1999) Estimating the hazard of rupture using uncertain occurrence times of paleoearthquakes. *J Geophys Res* 104:17,995–18,014
- Oglesby DD (1999) *Earthquake dynamics on dip-slip faults* [Thesis]. Type, University of California, Santa Barbara
- Oglesby DD, Archuleta RJ, Nielsen SB (1998) Earthquakes on dipping faults: the effects of broken symmetry. *Science* 280:1055–1059
- Okada Y (1985) Surface deformation due to shear and tensile faults in a half-space. *Bull Seismol Soc Am* 75:1135–1154
- Olsen KB, Archuleta RJ, Matarese JR (1995) Three-dimensional simulation of a magnitude 7.75 earthquake on the San Andreas fault. *Science* 270:1628–1632
- Parsons T (2002) Global Omori law decay of triggered earthquakes: large aftershocks outside the classical aftershock zone. *J Geophys Res* 107:2199. <https://doi.org/10.1029/2001JB000646>
- Parsons T (2008) Monte Carlo method for determining earthquake recurrence parameters from short paleoseismic catalogs: example calculations for California. *J Geophys Res* 113. <https://doi.org/10.1029/2007JB004998>
- Parsons T, Geist EL (2009) Tsunami probability in the Caribbean region. *Pure Appl Geophys* 165:2089–1226
- Peitgen H-O, Jürgens H, Saupe D (1992) *Chaos and fractals: new Frontiers of science*. Springer, New York
- Power W, Wang X, Lane E, Gillibrand P (2013) A probabilistic tsunami hazard study of the Auckland region, part I: propagation modelling and tsunami hazard assessment at the shoreline. *Pure Appl Geophys* 170:1621–1634
- Prevosto M, Kogstad HE, Robin A (2000) Probability distributions for maximum wave and crest heights. *Coast Eng* 40:329–360

- Rabinovich AB, Candella RN, Thomson RE (2011) Energy decay of the 2004 Sumatra tsunami in the World Ocean. *Pure Appl Geophys* 168:1919–1950
- Rayleigh JWS (1880) On the resultant of a large number of vibrations of the same pitch and of arbitrary phase. *Philos Mag* 10:73–78
- Ruiz JA, Fuentes M, Riquelme S, Campos J, Cisternas A (2015) Numerical simulation of tsunami runup in northern Chile based on non-uniform $k=2$ slip distributions. *Nat Hazards* 79:1177–1198
- Ryan KJ, Geist EL, Barall M, Oglesby DD (2015) Dynamic models of an earthquake and tsunami offshore Ventura, California. *Geophys Res Lett* 42:6599–6606
- Rychlik I, Leadbetter MR (1997) Analysis of ocean waves by crossing and oscillation intensities. In: *Proceedings of the seventh international offshore and polar engineering conference (ISOPE)*, vol 3, Golden, pp 206–213
- Rychlik I, Johannesson P, Leadbetter MR (1997) Modelling and statistical analysis of ocean-wave data using transformed Gaussian processes. *Mar Struct* 10:13–47
- Saito T, Furumura T (2009) Scattering of linear long-wave tsunamis due to randomly fluctuating sea-bottom topography: coda excitation and scattering attenuation. *Geophys J Int* 177:958–965
- Satake K (2007) Tsunamis. In: Kanamori H, Schubert G (eds) *Treatise on geophysics*, volume 4-earthquake seismology, vol 4. Elsevier, Amsterdam, pp 483–511
- Satake K, Kanamori H (1991) Use of tsunami waveforms for earthquake source study. *Nat Hazards* 4:193–208
- Satake K, Yoshida Y, Abe K (1992) Tsunami from the Mariana earthquake of April 5, 1990: its abnormal propagation and implications for tsunami potential from outer-rise earthquakes. *Geophys Res Lett* 19:301–304
- Satake K, Fujii Y, Harada T, Namegaya Y (2013) Time and space distribution of coseismic slip of the 2011 Tohoku earthquake as inferred from tsunami waveform data. *Bull Seismol Soc Am* 103:1473–1492
- Savage JC (1994) Empirical earthquake probabilities from observed recurrence intervals. *Bull Seismol Soc Am* 84:219–221
- Selva J, Tonini R, Molinari I, Tiberti MM, Romano F, Grezio A et al (2016) Quantification of source uncertainties in seismic probabilistic tsunami Hazard analysis (SPHA). *Geophys J Int* 205:1780–1803
- Sepúlveda I, Liu PL-F, Grigoriu M, Pritchard M (2017) Tsunami hazard assessments with consideration of uncertain earthquake slip distribution and location. *J Geophys Res Solid Earth* 122:7252–7271
- Sornette D, Sornette A (1999) General theory of the modified Gutenberg-Richter law for large seismic moments. *Bull Seismol Soc Am* 89:1121–1130
- Stark CP, Hovius N (2001) The characterization of landslide size distributions. *Geophys Res Lett* 28:1091–1094
- Strasser FO, Arango MC, Bommer JJ (2010) Scaling of the source dimensions of interface and intraslab subduction-zone earthquakes with moment magnitude. *Seismol Res Lett* 81:941–950
- Takahara M, Yomogida K (1992) Estimation of coda Q using the maximum likelihood method. *Pure Appl Geophys* 139:255–268
- Tamura S, Ide S (2011) Numerical study of splay faults in subduction zones: the effects of bimaterial interface and free surface. *J Geophys Res* 116. <https://doi.org/10.1029/2011JB008283>
- Tanioka Y, Satake K (1996) Tsunami generation by horizontal displacement of ocean bottom. *Geophys Res Lett* 23:861–865
- ten Brink US, Geist EL, Andrews BD (2006) Size distribution of submarine landslides and its implication to tsunami hazard in Puerto Rico. *Geophys Res Lett* 33. <https://doi.org/10.1029/2006GL026125>
- Titov VV, González FI, Bernard EN, Ebel JE, Mofjeld HO, Newman JC et al (2005) Real-time tsunami forecasting: challenges and solutions. *Nat Hazards* 35:40–58
- Tsai CP (1997) Slip, stress drop and ground motion of earthquakes: a view from the perspective of fractional Brownian motion. *Pure Appl Geophys* 149:689–706
- Turcotte DL (1992) *Fractals and chaos in geology and geophysics*. Cambridge University Press, Cambridge, p 221
- Turcotte D (2013) *Fractals in geology and geophysics*. In: Meyers R (ed) *Encyclopedia of complexity and systems science*. Springer, Berlin/Heidelberg
- van Dorn WG (1984) Some tsunami characteristic deducible from tide records. *J Phys Oceanogr* 14:353–363
- Vere-Jones D (2013) Earthquake occurrence and mechanisms, stochastic models for. In: Meyers R (ed) *Encyclopedia of complexity and systems science*. Springer, Berlin/Heidelberg
- Vere-Jones D, Robinson R, Yang W (2001) Remarks on the accelerated moment release model: problems of model formulation, simulation and estimation. *Geophys J Int* 144:517–531
- Wendt J, Oglesby DD, Geist EL (2009) Tsunamis and splay fault dynamics. *Geophys Res Lett* 36. <https://doi.org/10.1029/2009GL038295>
- Werner M, Sornette D (2013) Seismicity, statistical physics approaches to. In: Meyers R (ed) *Encyclopedia of complexity and systems science*. Springer, Berlin/Heidelberg
- Whirley RG, Engelmann BE (1993) *DYNA3D: a non-linear, explicit, three-dimensional finite element code for solid and structural mechanics – user manual*. University of California, Lawrence Livermore National Laboratory

Wedge Mechanics: Relation with Subduction Zone Earthquakes and Tsunamis

Kelin Wang^{1,2}, Yan Hu² and Jiangheng He¹

¹Pacific Geoscience Centre, Geological Survey of Canada, Sidney, Canada

²School of Earth and Ocean Sciences, University of Victoria, Victoria, Canada

Article Outline

Glossary

Definition of the Subject

Introduction

Stable and Critical Coulomb Wedges

Dynamic Coulomb Wedge

Stress Drop and Increase in a Subduction

Earthquake

Tsunamigenic Coseismic Seafloor Deformation

Future Directions

Bibliography

Glossary

Accretionary wedge (prism) At some subduction zones, as one plate subducts beneath the other, some sediment is scraped off the incoming plate and accreted to the leading edge of the upper plate. Because of its wedge shape, the accreted sedimentary body is called the accretionary wedge (or accretionary prism). If all the sediment on the incoming plate is subducted, there is still a sedimentary wedge in the frontal part of the upper plate, but it is usually very small and consists of sediments derived from the upper plate by surface erosion.

Coulomb plasticity Coulomb plasticity is a macroscopic, continuum description of the most common type of permanent deformation of Earth materials such as sand, soil, and rock at relatively low temperature and pressure and

is widely used in civil engineering and Earth science. In detail, the deformation mechanism is actually brittle shear failure, with or without emitting elastic wave energy. The macroscopic yield criterion is the Coulomb's law, in which shear strength increases linearly with confining pressure. If the strength does not change with permanent deformation, the material is said to be perfectly plastic. Note that in Earth science the word plasticity is also used to indicate thermally activated creep, but it is very different from the meaning used herein.

Subduction zone earthquake cycle Megathrust fault, the interface between the two converging lithospheric plates at a subduction zone, moves in a stick-slip fashion. In the "stick" phase, the fault is locked or slips very slowly, allowing elastic strain energy to be accumulated in both plates around the fault. Every few decades or centuries, the fault breaks as high-rate "slip" to release the strain energy, causing a large or great earthquake, usually accompanied with a tsunami. An interseismic period and the ensuing earthquake together is called a subduction zone earthquake cycle. The word cycle by no means implies periodicity. Neighboring segments of the same subduction zone may exhibit different temporal patterns of earthquake cycles.

Velocity-weakening and strengthening These are macroscopic descriptions of dynamic frictional behavior of contact surfaces. Velocity-weakening, featuring a net decrease in frictional strength with increasing slip rate, is the necessary condition for a fault to produce earthquakes. It differs from slip-weakening in that a velocity-weakened fault will regain its strength when the slip slows down or stops. Velocity-strengthening is the opposite of velocity-weakening and is the necessary condition for a fault to resist earthquake rupture. Detailed physical processes on the contact surfaces or within the fault zones controlling their frictional behavior are still being investigated.

Definition of the Subject

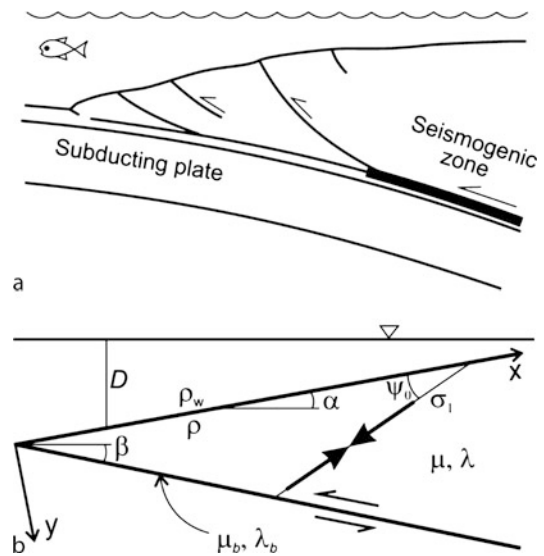
Mechanics of wedge-shaped geological bodies such as accretionary prisms at subduction zones and fold-and-thrust belts at collision zones is of great scientific interest, mainly because it enables us to use the observed morphology and deformation of the wedge-shaped bodies to constrain properties of the thrust faults underlying them. Davis et al. (1983) drew analogy between these wedge-shaped bodies and the sand wedge in front of a moving bulldozer and established a mathematical model. Their work triggered wide applications of wedge mechanics to geology. The fundamental process described in wedge mechanics is how gravitational force, in the presence of a sloping surface, is balanced by basal stress and internal stress. The internal state of stress depends on the rheology of the wedge. The most commonly assumed wedge rheology for geological problems is perfect Coulomb plasticity (Davis et al. 1983), and the model based on this rheology is referred to as the Coulomb wedge model.

The Coulomb wedge model was designed to explain geological processes of timescale of hundreds of thousands of years. The state of stress is understood to be an average over time, and the wedge is assumed to be in a critical state, that is, uniformly at the Coulomb yield stress. In the application of the model to the sedimentary wedges at subduction zones, attention is now being paid to the temporal changes in stress and pore fluid pressure associated with great subduction earthquakes which have recurrence intervals of decades to centuries. To account for the shortterm stress changes, Wang and Hu (2006) expanded the Coulomb wedge model by introducing the elastic – perfectly Coulomb plastic rheology. The expanded Coulomb wedge model links long-term geology with coseismic processes and provides a new perspective for the study of subduction zone earthquakes, tsunami generation, frontal wedge structure, and forearc hydrogeology.

Introduction

Sloping surfaces are commonly dealt with in engineering problems such as dam design and

landslide hazard mitigation. In the presence of a sloping surface, materials under gravitational force tend to “flow” downhill and thus generate tensile stress, but whether collapse actually occurs depends on the strength of the material. Materials used in construction can easily sustain stresses caused by the presence of a vertical surface, but a material of no shear strength such as stationary water cannot support any surface slope. A geological wedge, such as the accretionary prism at a subduction zone (see Fig. 1a), overlies a dipping fault, and hence its internal stress is controlled by the strength of the fault as well. If the basal fault is a thrust fault and is strong relative to the strength of the wedge material, the wedge can undergo compressive failure. Wedge mechanics thus consists of two aspects: the frictional behavior of the basal fault and the deformation of the wedge itself. Given a wedge with density ρ , surface slope angle α , and basal dip β (see Fig. 1b), Elliot (1976), Chapple (1978), and later



Wedge Mechanics: Relation with Subduction Zone Earthquakes and Tsunamis, Fig. 1 (a) Schematic illustration of a subduction zone accretionary wedge. (b) Coordinate system used in this article (x, y). The surface of the wedge and the subduction fault are simplified to be planar, with slope angle α and dip β , respectively, and the x axis is aligned with the upper surface. ψ_0 is the angle between the maximum compressive stress σ_1 and the upper surface. D is water depth. ρ and ρ_w are densities of the wedge material and overlying water, respectively. μ , λ , μ_b and λ_b are as defined in Eqs. (2) and (3)

workers all considered the following equations of force balance (exact form varies between publications depending on the assumed coordinate systems)

$$\frac{\partial \sigma_x}{\partial x} + \frac{\partial \tau_{xy}}{\partial y} - \rho g \sin \alpha = 0, \quad (1a)$$

$$\frac{\partial \tau_{xy}}{\partial x} + \frac{\partial \sigma_y}{\partial y} + \rho g \cos \alpha = 0, \quad (1b)$$

where g is gravitational acceleration, σ_x and σ_y are normal stresses, and τ_{xy} is shear stress.

It was Davis et al. (1983) who introduced Coulomb plasticity into wedge mechanics. Coulomb plasticity was initially proposed by French engineer Coulomb in 1773 to describe the mechanical strength of soil and sand. The Coulomb failure criterion in its simplest form is

$$\tau = S - \mu(\sigma_n + P) = S - \mu \bar{\sigma}_n, \quad (2)$$

where τ is shear strength, σ_n is normal stress, S is cohesion, and $\mu = \tan \varphi$ is the coefficient of internal friction with φ called the internal friction angle. P is the pressure of fluids present in the pore space between solid grains and in various small fractures and is loosely referred to as the pore fluid pressure. Note that the effective stress $\bar{\sigma}_n = \sigma_n + P$ is normal stress with P subtracted. The plus sign is due to the custom in mechanics (except rock mechanics) that compressive stress is defined to be negative, but pressure, although also compressive, is defined to be positive. Here the plane on which the stress is evaluated is oriented in any arbitrary direction, but failure will start on the set of planes that meets the above criterion. This is a generalization of the Coulomb friction criterion for a fixed fault plane

$$\tau_b = S_b - \mu_b(\sigma_{n_b} + P_b) = S_b - \mu_b \bar{\sigma}_{n_b}, \quad (3)$$

where S_b is the cohesion of the fault, $\mu_b = \tan \varphi_b$ is its friction coefficient, and P_b is fluid pressure along the fault. S_b is usually negligibly small and almost always taken to be zero, and μ_b is normally significantly lower than μ . A well developed fault such as a plate boundary fault is a zone of finite thickness filled with gouge material, so that the

“friction” described by (3) or other friction laws is actually the shear deformation of the gouge in the fault zone, and P_b is actually pore fluid pressure of the gouge. Fault gouge is often made very weak by the presence of hydrous minerals (Brown et al. 2003; Takahashi et al. 2007), such that the collective strength of the fault zone material is much less than the strength of the rocks on both sides. Another process that may weaken the fault is that the local hydrogeological regime may dynamically maintain P_b in the fault zone to stay higher than P on both sides (Dewhurst et al. 1996). For both Coulomb plasticity and Coulomb friction, strength increases with depth because of the increasing pressure thus normal stress.

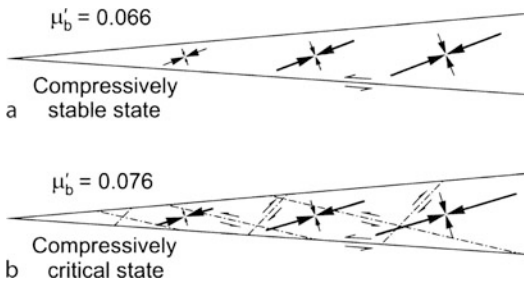
It has long been recognized that Coulomb plasticity, featuring strong depth dependence, applies to the shallow part of Earth’s lithosphere. The most common example is the use of Byerlee’s law of rock friction (Byerlee 1978) to describe brittle strength in “Christmas-tree”-like vertical strength-depth profiles of the lithosphere. The Byerlee’s law is an empirical Coulomb friction law. By assuming that faults, i.e., potential failure planes, are oriented in all directions, we regard the brittle part of the lithosphere as being Coulomb plastic. This example also illustrates how a system of numerous discrete structures can be regarded as a continuum at a much larger scale. Similarly, although a geological wedge actually consists of numerous blocks divided by fractures, Coulomb plasticity can be used to describe its overall rheology. However, specific values of friction parameters for submarine wedges may be quite different from those in the Byerlee’s law.

The Coulomb wedge model explains how the geometry (taper) of the wedge is controlled by the interplay between the gravitational force, the strength of the wedge material, and the strength of the basal fault. The wedge strength and fault strength are both strongly influenced by pore fluid pressure, and the most popular application of the model is to estimate pore fluid pressure from observed wedge geometry. Since the work of Davis et al. (1983), more rigorous analytical solutions have been derived (Dahlen 1984; Dahlen et al. 1984; Fletcher 1989; Wang and Hu 2006; Zhao et al. 1986), and some extensions have been proposed, e.g., (Barr and Dahlen 1990; Breen and

Orange 1992; Enlow and Koons 1998; Xiao et al. 1991). Lithospheric scale numerical models are often used to study the evolution of geological wedges including such effects as erosion and sedimentation in collision or subduction zone settings, e.g., (Fuller et al. 2006; Willett et al. 1993). Utilizing the bulldozer – sand wedge analogy, important physical insights have been obtained from sandbox experiments (Byrne et al. 1988; Davis 1990; Kukowski et al. 1994; Lallemand et al. 1994; Lohrmann et al. 2003; Mourgues and Cobbold 2006; Smit et al. 2003; Wang and Davis 1996). For a list of applications of the Coulomb wedge model to subduction-zone accretionary prisms, see Wang and Hu (2006).

Stable and Critical Coulomb Wedges

Depending on the state of stress, a Coulomb wedge can be at a critical state, that is, everywhere at Coulomb failure, or a stable (also referred to as supercritical) state, that is, everywhere not at failure (see Fig. 2). The taper of a critical wedge will not change if the stress does not change; note that a critical wedge of stable geometry should not be confused with a stable wedge. Stress solutions have been derived for both critical and stable states. Here we only summarize the simplest, exact solution for a cohesionless wedge ($S = 0$)



Wedge Mechanics: Relation with Subduction Zone Earthquakes and Tsunamis, Fig. 2 An example to show how stresses in an elastic – perfectly Coulomb-plastic wedge, with $\alpha = 5^\circ$, $\beta = 4^\circ$, $\mu = 0.6$, and $\lambda = 0.86$, are affected by basal friction $\mu'_b = \mu_b(1 - \lambda_b)$. Converging arrows represent principal stresses, with the larger one being σ_1 . (a) Compressively stable state. (b) Compressively critical state. Dot-dashed lines are plastic slip lines (potential failure planes)

derived in the coordinate system shown in Fig. 1b, because of the convenience of its application as compared to other solutions. The wedge is assumed to be elastic – perfectly Coulomb plastic. If it is at the critical state, it obeys (2); if it is at the stable state, it obeys the Hooke's law of elasticity. The basal thrust fault obeys (3). The wedge is assumed to be under water of density ρ_w and depth D (a function of x ; see Fig. 1b). By defining a Hubbert–Rubey fluid pressure ratio within the wedge (Dahlen 1984)

$$\lambda = \frac{P - \rho_w g D}{-\sigma_y - \rho_w g D}, \quad (4)$$

and a similar parameter along the basal fault (Wang et al. 2006)

$$\lambda_b = \frac{P_b - \rho_w g D + \lambda H}{-\sigma_y - \rho_w g D + H}, \quad (5a)$$

where

$$H = \frac{\sigma_y - \sigma_n}{1 - \lambda}, \quad (5b)$$

and assuming $S_b = 0$, we can rewrite (3) into

$$\tau_b = -\mu_b \bar{\sigma}_n = -\mu''_b \bar{\sigma}_n, \quad (6a)$$

where

$$\mu''_b = \frac{1 - \lambda_b}{1 - \lambda} \mu_b = \frac{\mu'_b}{1 - \lambda}. \quad (6b)$$

The strength of the basal fault is represented by $\mu'_b = \mu_b(1 - \lambda_b)$ which always appears as a single parameter and is commonly referred to as the effective friction coefficient. The hydrological process in the fault zone may differ from that in the wedge and thus cause a sharp gradient in fluid pressure across the wedge base. By allowing λ_b to be different from λ , we use a pressure discontinuity to approximate the sharp gradient. Because stress is continuous across the basal fault, the discontinuity in pore fluid pressure leads to a discontinuity in effective stress. The second equality of Eq. (6a) shows the relation between

the effective stress along the fault ($\bar{\sigma}_{n_b}$) and the effective stress just above the fault ($\bar{\sigma}_n$). The establishment of (6) allows the following exact stress solution to be derived. In this expression, all stress components are normalized by ρgy (e.g., $\bar{\sigma}'_x = \bar{\sigma}_x / \rho gy$).

$$\bar{\sigma}'_x = -m(1 - \lambda) \cos \alpha, \quad (7a)$$

$$\bar{\sigma}'_y = -(1 - \lambda) \cos \alpha, \quad (7b)$$

$$\tau'_{xy} = (1 - \rho') \sin \alpha, \quad (7c)$$

where $\rho' = \rho_w / \rho$, and the effective stress ratio $m = \bar{\sigma}_x / \bar{\sigma}_y$ depends on whether the wedge is stable or critical. If the wedge is in a stable state (elastic) (Wang and Hu 2006),

$$m = 1 + \frac{2(\tan \alpha' + \mu_b'')}{\sin 2\theta(1 - \mu_b'' \tan \theta)} - \frac{2 \tan \alpha'}{\tan \theta}, \quad (8)$$

where $\theta = \alpha + \beta$, and,

$$\tan \alpha' = \frac{1 - \rho'}{1 - \lambda} \tan \alpha. \quad (9)$$

The angle ψ_0 between the most compressive stress σ_1 and the upper surface is uniform (see Fig. 1b). A more general solution for a purely elastic wedge can be found in Hu and Wang (2006). If the wedge is in a critical state (perfectly plastic) (Dahlen 1984)

$$m = m^c = 1 + \frac{2 \tan \alpha'}{\tan 2\psi_0^c}, \quad (10)$$

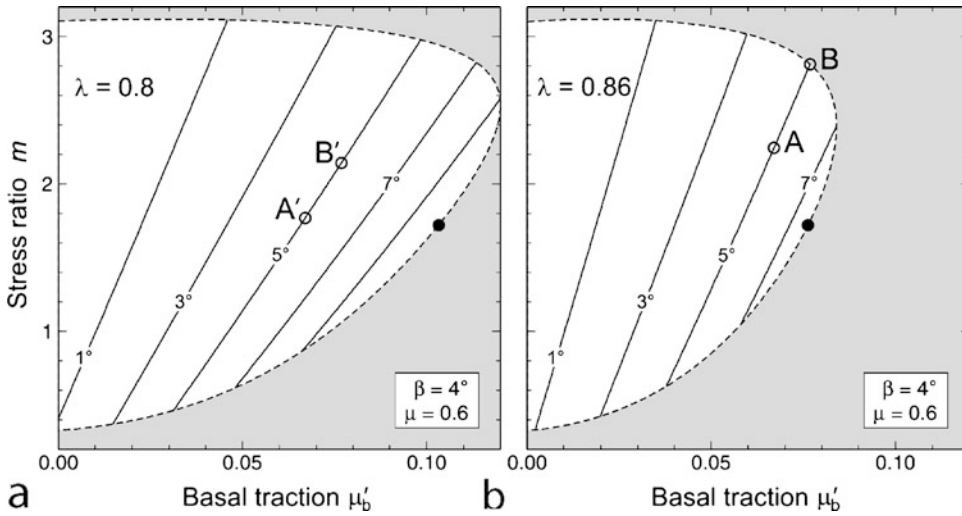
where ψ_0^c is the value of ψ_0 in the critical state and is given by the following relation

$$\frac{\sin \varphi \sin 2\psi_0^c}{1 - \sin \varphi \cos 2\psi_0^c} = \tan \alpha'. \quad (11)$$

In the above expressions, superscript c indicates critical state. If the wedge has a cohesion that is proportional to depth, (10) and (11) will be modified only slightly (Wang and Hu 2006; Zhao et al. 1986).

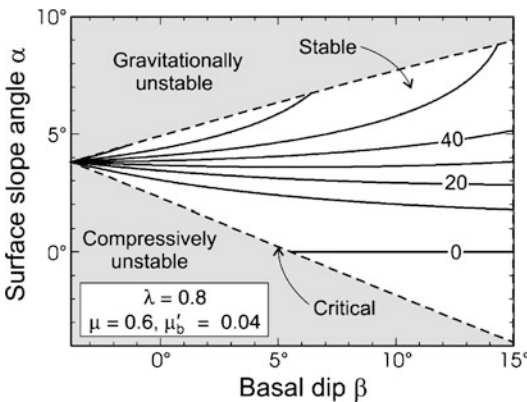
A wedge of fixed geometry has two m^c values. The lower one defines the extensionally critical state, in which the wedge is on the verge of gravitational collapse. This occurs if friction along the basal fault is very low relative to the strength of the wedge material. The higher one defines the compressively critical state, in which the wedge is everywhere on the verge of thrust failure. If m lies between these two critical values, the wedge is in a stable state and only experiences elastic deformation. A change of basal friction μ_b' will cause a change in m and thus may potentially cause the wedge to switch between the stable and critical states (see Fig. 3). An example of a wedge being in a stable or compressively critical state as controlled by basal friction is shown in Fig. 2. The plastic slip lines (potential failure planes) in the critical wedge (see Fig. 2b) are reminiscent of the out-of-sequence faults in a real accretionary prism (see Fig. 1a).

If we fix the values of all material properties, the critical-wedge solution ($m = m^c$) defines a relation between α and β representing all possible geometries of a critically tapered wedge (dashed line in Fig. 4). This is a very commonly used diagram, in which the lower branch of the $\alpha - \beta$ curve represents compressively critical states, and the upper branch represents extensionally critical states. Combinations of α and β outside of the stability region comprise unstable geometries and cannot exist in steady state. If sedimentary wedges of subduction zones are compressively critical, their observed $\alpha - \beta$ pairs should line up with the lower branch. However, it has been shown that most of them fall in the stable or even extensionally unstable region of this type of diagram (Lallemand et al. 1994; von Huene and Ranero 2003). To fit observations, we need to move the lower branch upward by a significant amount. In order to do this, we need to assume either a weaker wedge or a stronger basal fault, or both. This is more simply illustrated by Fig. 3. Given wedge geometry and internal friction, if the state of the wedge is to be changed from stable to compressively critical, we need to have a higher friction (μ_b') along the basal fault (i.e., greater stress coupling) and/or higher pore fluid pressure within the wedge (i.e., greater λ weakening the



Wedge Mechanics: Relation with Subduction Zone Earthquakes and Tsunamis, Fig. 3 Effective stress ratio m (Eq. (7)) as a function of basal friction $\mu'_b = \mu_b(1 - \lambda_b)$ for wedges of the same basal dip ($\beta = 4^\circ$) but different surface slope angles (α) as labelled on the curves (solid lines). (a) and (b) are for two different pore fluid pressure ratio values within the wedge. Each curve is terminated at the extensionally critical state at a lower μ'_b and the compressively critical state at a higher μ'_b . The end points (connected by a dashed line) outline the stable region (white). No solution exists outside this region. The

solid circle marks the state in which the surface slope is at the angle of repose. It divides the line of critical states (dashed line) into the compressive part (above) and extensional part (below). Open circles in (b) labelled A and B mark the states shown in Fig. 2a, b, respectively. State A' in (a) is for the same wedge with the same basal friction as state A in (b) except for a lower pore fluid pressure ratio, and state B' in (a) corresponds to state B in (b) in the same fashion. Comparison of B with B' shows how an increase in pore fluid pressure weakens the wedge

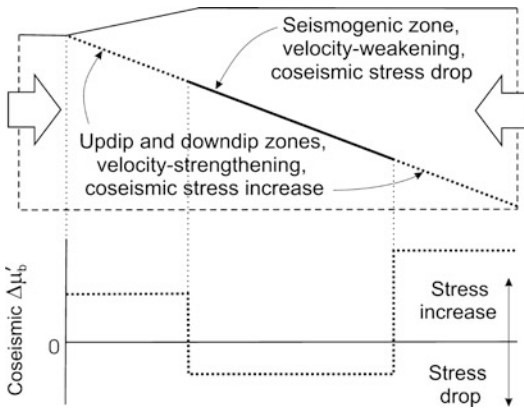


Wedge Mechanics: Relation with Subduction Zone Earthquakes and Tsunamis, Fig. 4 Possible wedge geometry (α and β) given material strength and basal friction. The dashed line indicates critical wedge geometry, with the upper and lower branches representing extensionally and compressively critical states, respectively. Contours of ψ_0 (in degrees) are shown in the stable region (white)

wedge material by reducing effective pressure). Wang and Hu (2006) proposed that higher μ'_b and λ can occur at the time of a great earthquake and introduced the concept of dynamic Coulomb wedge.

Dynamic Coulomb Wedge

The concept of dynamic Coulomb wedge is based on the widely recognized frictional behavior of subduction faults. Ignoring the presence of along-strike variations of frictional properties, we can summarize the frictional behavior in a simplified cross-section view (see Fig. 5). The seismogenic zone exhibits velocity-weakening behavior: It weakens in response to high-rate slip, resulting in slip instability, that is, earthquakes. The segments updip and downdip of the seismogenic zone exhibits velocitystrengthening: They strengthen at the time of the earthquake to develop



Wedge Mechanics: Relation with Subduction Zone Earthquakes and Tsunamis, Fig. 5 Schematic illustration of the subduction zone model considered in this work.

Large arrows represent interseismic strain accumulation. An earthquake is represented by a sudden decrease in the effective friction coefficient μ'_b of the seismogenic zone by $\Delta\mu'_b$. Coseismic strengthening of the updip and downdip zones is represented by a sudden increase in their μ'_b values

higher stress to resist rupture, but they may slip aseismically after the earthquake to relieve the high stress attained during the earthquake. We assume that the actively deforming sedimentary wedge overlies the updip segment (also see Fig. 1a).

Microscopic mechanisms for the velocity-weakening behavior of the seismogenic zone and the velocitystrengthening behavior of the aseismic zones are subjects of intense research. The aseismic behavior of the deeper part of any fault is intuitively easy to comprehend; higher temperature at greater depths increasingly enhances viscous deformation and inhibits brittle faulting. There are different physical explanations for the velocity-weakening behavior of the seismogenic zone as summarized in Rice (2006). The mechanism responsible for the velocity-strengthening behavior of the updip segment is yet to be identified, although it is widely accepted that the presence of clay minerals has something to do with it (Hyndman and Wang 1993; Moore and Lockner 2007; Moore and Saffer 2001; Vrolijk 1990; Wang 1980). Laboratory experiments indicate that dilatancy of granular fault zone material during fast slip can lead to velocity-strengthening (Morone 1998). We think part of the reason for

the velocitystrengthening behavior of the updip segment may be its inability to localize into a very thin slip zone. Seismic rupture occurs along very thin slip zones of a few millimeters thickness that are parts of a thicker fault zone (Sibson 2003). Fast slip of the updip segment, if triggered by the rupture of the deeper seismogenic zone, may tend to drag along fault zone materials over a more distributed band and thus meet greater resistance. This view is different from the velocity-strengthening process described by laboratory-derived rate- and state-dependent friction laws (Dieterich 1979; Ruina 1983) in which dynamic changes in the thickness of the slip zone plays no role.

Regardless of the microscopic mechanisms, the downdip variation of the frictional behavior is expected to bring direct consequences to wedge deformation. In an earthquake, thrust motion of the seismogenic zone causes the frontal wedge to be pushed from behind, and velocity-strengthening of the updip megathrust segment gives rise to higher stress at its base. If the wedge is originally in a stable state, the coseismic strengthening of the basal fault may increase the m value in (7a) from subcritical to critical. After the earthquake, when the seismogenic zone has returned to a locked state, the stress along the updip segment will relax. The decrease in μ'_b leads to a smaller m , and therefore the frontal wedge returns to a stable state.

If we ignore the change in fluid pressure, the above described process can be seen as the stress ratio m moving up-and-down along one of the solid lines in Fig. 3 in response to changes in μ'_b in earthquake cycles. During a big earthquake, it will hit the upper end of the line (m^c). The state of stress for a stable wedge before the earthquake ($m < m^c$) and that for a compressively critical wedge during the earthquake ($m = m^c$) are illustrated by examples in Fig. 2a, b, respectively. Fluid pressure variation should not be ignored, however. For example, the same basal friction as shown in Fig. 2b (also see point B in Fig. 3b) will not drive the wedge to failure if the pore fluid pressure is lower, as shown by point B⁰ in Fig. 3a.

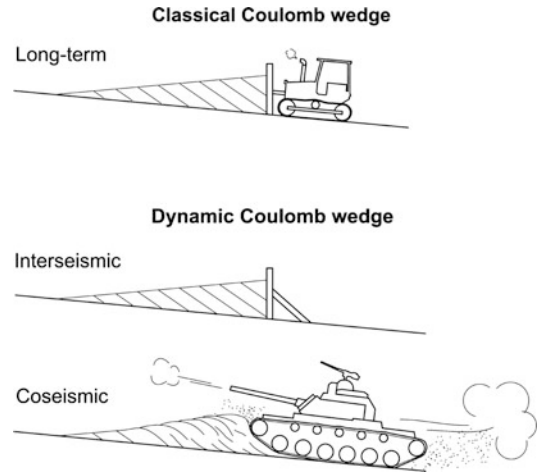
During an earthquake, the sudden compression of the frontal wedge will cause its internal pore

fluid pressure to increase, coseismically weakening the wedge material. By comparing Fig. 3a, b, we can see that if the pore fluid pressure ratio in the wedge is higher, the increase in basal friction required to push the wedge into a critical state can be smaller. We may envisage the following scenario. The pore fluid pressure in a frontal wedge may decrease to some degree over the interseismic period due to fluid drainage through fractures and stress relaxation, and the mechanical state of the wedge before an earthquake can be represented by point A' in Fig. 3a as opposed to point A in Fig. 3b. An earthquake not only causes the basal friction to increase but also the pore fluid pressure within the wedge to rise, such that the wedge enters a critical state represented by point B in Fig. 3b. Therefore, the coseismic strengthening of the basal fault and coseismic weakening of the wedge both work toward bringing the wedge to failure.

All previous applications of the Coulomb wedge model to subduction zones assume $m = m^c$. The dynamic Coulomb wedge model of (Wang and Hu 2006) explains the meaning of this long-term m^c : At least as an end-member scenario, it is the value of m briefly achieved in numerous large earthquakes. The “average” basal stress that determines the wedge geometry in long-term Coulomb wedge models is actually the peak stress achieved at the time of large earthquakes. Thus, the common illustration of the peaceful scene of a bulldozer pushing a sand wedge in classical Coulomb wedge papers (see Fig. 6a) should be modified to reflect the unpleasant reality of the world (see Fig. 6b).

Stress Drop and Increase in a Subduction Earthquake

It is important to know the possible amount of stress increase in the frontal wedge for a given earthquake. The increase cannot be arbitrarily large; it is limited by the level of the “push” on the wedge from behind during an earthquake. For this purpose, a numerical model of a larger scale embracing the essential components of the subduction fault as shown in Fig. 5 must be



Wedge Mechanics: Relation with Subduction Zone Earthquakes and Tsunamis, Fig. 6 Cartoon illustrating the difference between the classical Coulomb wedge model and the dynamic Coulomb wedge model for subduction zone accretionary prisms. In the classical wedge model, $m = m^c$. In the dynamic wedge model, $m < m^c$ in the interseismic period but $m = m^c$ at the time of a large earthquake. See Eqs. (7) and (10) for definition of m and m^c

considered, because the stress interaction between the frontal wedge and the material overlying the seismogenic zone cannot be handled by the analytical Coulomb wedge solutions. For an illustration, we consider the following model geometry, representative of most subduction zones. The subduction fault has a constant dip $\beta = 4^\circ$. The frontal 50 km of the upper plate has a surface slope $\alpha = 5^\circ$, representing the sedimentary wedge, and the rest of the upper plate has a flat surface. We wish to focus on the process of stress transfer from the seismogenic zone to the updip segment during an earthquake, and a static, uniform, and purely elastic model suffices. For simplicity, the effect of pore fluid pressure change on deformation is neglected.

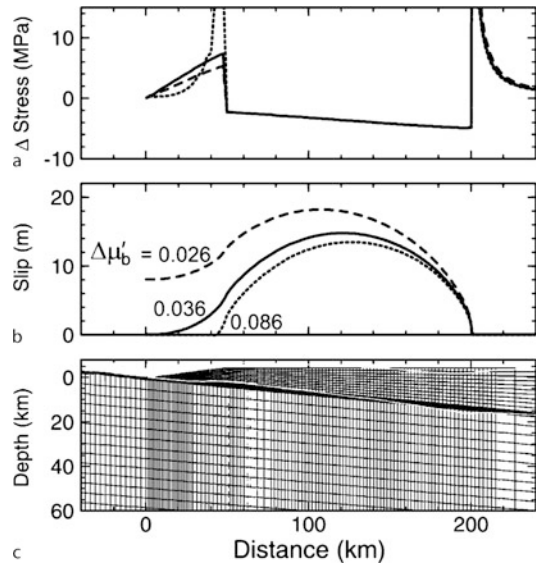
We use a 2D plane-strain finite element model and simulate Coulomb friction along the fault using the method of Lagrange-multiplier Domain Decomposition (Wang and He 1999). The model boundaries are set sufficiently far away so that the model resembles a half-space. For numerical stability, we invoke gravity (assuming a rock density of 2800 kg/m^3 only when determining yield stress along the fault but exclude it from the deformation

calculation. The effect of gravity on coseismic elastic deformation is very small and is neglected in most earthquake cycle deformation models, but gravity is important in the calculation of frictional slip of the fault.

We first generate a pre-stress field by moving the remote seaward and landward model boundaries toward each other against a locked fault. At this stage of “interseismic” strain accumulation, we use a μ'_b of 0.04 for the seismogenic and updip segments and 0.004 for the deeper segment. The low strength of the subduction fault is based on the weak fault argument as summarized in Wang and Hu (2006). The nearly zero strength of the deeper part represents a relaxed state after a long time of locking of the seismogenic zone. However, the absolute strength of the fault has no effect on our results. It is the incremental change in fault strength ($\Delta\mu'_b$) at the time of the earthquake that is relevant. A negative $\Delta\mu'_b$ represents the net effect of weakening, and a positive $\Delta\mu'_b$ represents the net effect of strengthening. The velocity-dependent evolution of $\Delta\mu'_b$ through time is not explicitly simulated.

Three examples are shown in Fig. 7. In all cases, $\Delta\mu'_b = -0.01$ is assigned to the 150-km wide seismogenic zone. This value is chosen to produce an average stress drop of a few MPa (see Fig. 7a), typical of values observed for great subduction earthquakes. The stress drop releases elastic strain energy initially stored in the system, leading to fault slip that represents an earthquake rupture. The deepest segment is assigned a sufficiently large positive $\Delta\mu'_b$ so that it cannot slip. The examples differ in the $\Delta\mu'_b$ values assumed for their 50 km wide updip segment, which is the coseismic increase in basal friction in the dynamic Coulomb wedge model.

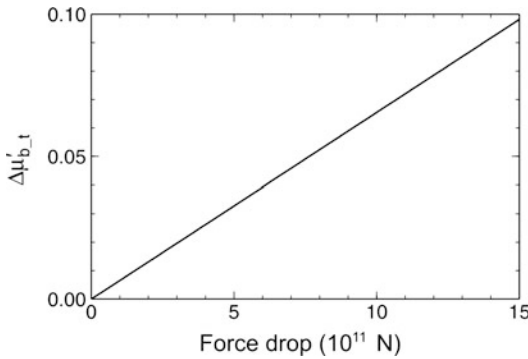
Example 1 No trench-breaking rupture (solid line in Fig. 7b). In this case, the strengthening of the updip segment is $\Delta\mu'_b = 0.036$. This particular value of $\Delta\mu'_b$ creates a situation in which the entire updip segment is at failure but is just short of breaking the trench. This is the minimum value of the $\Delta\mu'_b$ of the updip segment required to prevent trench-breaking rupture and is denoted $\Delta\mu'_{b_t}$. The value of $\Delta\mu'_{b_t}$ depends on the product



Wedge Mechanics: Relation with Subduction Zone Earthquakes and Tsunamis, Fig. 7 Three examples of the stress transfer model. The examples differ in the coseismic strength increase of the updip fault segment, indicated in (b) as $\Delta\mu'_b$. (a) Shear stress drop (or increase) along the fault. (b) Slip distribution along the fault. (c) Central portion of the finite element mesh (thin lines). The “thick line” along the plate interface is actually a group of very densely spaced elements. Thick gray lines indicate deformed fault and surface after the earthquake (exaggerated by a factor of 2000)

of the stress drop and the area of the seismogenic zone, a quantity we refer to as “force drop”. That is, if the upper edge of the seismogenic zone is fixed, increasing its downdip width or stress drop gives the same result. For the same model geometry as used for this example, $\Delta\mu'_{b_t}$ as a function of force drop per unit strike length is shown in Fig. 8, assuming the upper edge of the seismogenic zone is fixed. Using a different model geometry or position of the seismogenic zone upper edge will change the slope of this function.

Example 2 Trench-breaking rupture (dashed line in Fig. 7b). Given the same force drop in the seismogenic zone, if $\Delta\mu'_b$ of the updip segment is less than $\Delta\mu'_{b_t}$, the rupture will break the trench. Knowing whether coseismic trench-breaking rupture exists or is common awaits future seafloor monitoring observations. This example shows



Wedge Mechanics: Relation with Subduction Zone Earthquakes and Tsunamis, Fig. 8 Minimum increase in $\Delta\mu'_b$ of the updip megathrust segment (denoted $\Delta\mu'_{b,t}$) required to prevent trench-breaking rupture as a function of force drop along the seismogenic zone for the model shown in Fig. 7c, with the upper edge of the seismogenic zone fixed at 50 km from the trench

that a trench-breaking rupture does not necessarily indicate the updip segment exhibits velocity-weakening. Conceivably, the slip of the velocity-strengthening updip segment may be slower than that of the seismogenic zone and may not generate much seismic waves.

Example 3 Fully buried rupture (dotted line in Fig. 7b). If $\Delta\mu'_b$ of the updip segment is greater than $\Delta\mu'_{b,t}$, rupture may only extend into its lower part. For a very high $\Delta\mu'_b$, most of the segment does not slip at all, because a tiny portion immediately updip of the seismogenic zone is sufficient to stop the rupture. This is just the buried-rupture scenario of the crack model commonly used in earthquake simulation (Geist and Dmowska 1999). Because most of the updip segment is “protected” and does not experience coseismic stress increase, this scenario is not applicable to the dynamic Coulomb wedge model. A very large stress increase just updip of the seismogenic zone is considered unrealistic.

These examples show the consequences of changes in the strength of the basal fault of the frontal wedge resulting from the rupture of the seismogenic zone. Whether the given strength increase $\Delta\mu'_b$ can drive the wedge from a stable state into a critical state depends on two factors.

First, it depends on the value of $\Delta\mu'_b$ before the earthquake. The value of 0.04 used in the above examples is only one of the numerous possible values. If $\Delta\mu'_b$ is already near a critical value, that is, m in Fig. 3 for a given α is already near m^c , a small increase will do. Conceivably, μ'_b before an earthquake may be relatively high if the strengthened state of the fault caused by the previous earthquake has not fully relaxed. Second, given μ'_b , it depends on the strength of the wedge material. A weaker wedge becomes critical at a lower $\Delta\mu'_b$. The average internal friction value μ of an actively deforming frontal prism is lower than the rest of the lithosphere because of its low degree of consolidation and high degree of fracturing. The pore fluid pressure within it, represented by λ in the Coulomb wedge model, may increase due to coseismic compression of the prism, further weakening the wedge, as discussed in section “Dynamic Coulomb Wedge”.

Tsunamigenic Coseismic Seafloor Deformation

To understand the process of tsunami generation by a great subduction zone earthquake, we must know how the seafloor deforms at the time of the earthquake. Despite the dramatic worldwide improvement of geodetic, seismo-logical, and oceanographic monitoring networks over the past few decades, our knowledge of coseismic seafloor deformation (CSD) is surprisingly poor and is based almost entirely on theoretical models. The problem is the rarity of near-field observations. Except for seafloor pressure sensor records at the time of the M8.2 Tokachi-oki earthquake of 2003 (Baba et al. 2006) and continuous GPS measurements from islands very near the Sumatra trench at the time of the M8.7 Nias-Simeulue earthquake of 2005 (Briggs et al. 2006; Hsu et al. 2006), most observations are made at sites too remotely located to resolve reliably CSD within about 100 km of the trench. The lack of near-field CSD information causes severe non-uniqueness in the inversion of tsunami, seismic, and geodetic data to determine coseismic slip patterns of the shallow part of the subduction

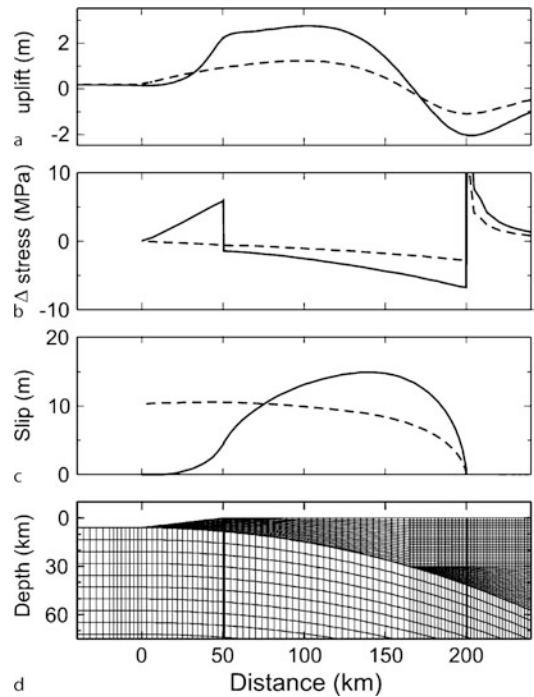
fault, for which the only cure is to introduce a priori constraints on the basis of theoretical models. Until the situation is improved by the establishment of seafloor monitoring systems, we must continue to resort to what we are able to deduce from these models.

The largest uncertainty in our knowledge of the processes that control CSD is how coseismic slip along the subduction fault varies in the downdip direction (Geist and Dmowska 1999; Wang and He 2007). It may overshadow uncertainties in our knowledge of the timescale of the deformation and the spatial variation of mechanical properties of the rock medium. In comparison, along-strike variations of the coseismic slip, usually described in terms of “asperities”, are much easier to determine using high-density terrestrial monitoring networks. Theoretical models discussed above can help us understand the downdip slip distribution. Using the same type of model as shown in Figs. 5 and 7 but with a realistic, curved fault, we illustrate how the frictional behavior of the updip segment affects the CSD (see Fig. 9). The simulated earthquakes in the two shown examples have the same “size”, quantified by the seismic moment – the product of rigidity, slip area, and average slip, but they cause very different CSD patterns.

Example 4 No trench-breaking rupture (solid line in Fig. 9a). This model is similar to the first example in the preceding section in that the updip segment is assumed to strengthen by $\Delta\mu'_{b-}$, and the rupture is on the verge of breaking the trench.

Example 5 Full trench-breaking rupture (dashed line in Fig. 9a). In this model, there is no velocity-strengthening updip segment, and the seismogenic zone extends to the trench. The stress drop of the resultant much wider seismogenic zone features a monotonic increase from the trench.

The model of full trench-breaking rupture yields much smaller vertical CSD than does the model of no trenchbreaking rupture. The reason is two-fold. First, without a velocity-strengthening updip segment to resist rupture, the maximum slip



Wedge Mechanics: Relation with Subduction Zone Earthquakes and Tsunamis, Fig. 9 Two examples showing how the frictional behavior of the updip segment (see Fig. 5) affects CSD. In one example (solid line), the segment strengthens by $\Delta\mu'_{b-}$, and in the other example (dashed line), the segment weakens and thus becomes part of the seismogenic zone. The simulated earthquakes in both examples have the same seismic moment. (a) Surface uplift. (b) Stress drop (or increase) along the fault. (c) Slip distribution along the fault. (d) Central part of the finite element mesh (thin lines). The two vertical “thick lines” at distances 50 km and 200 km bracketing the seismogenic zone of the model of no trench-breaking rupture are actually groups of very densely spaced elements

occurs in the most shallowly dipping near-trench part of the fault where seafloor displacement is predominantly horizontal. This effect would not be obvious had a straight fault geometry been used in the model. Second, without the resistance of an updip segment, the upper plate does not experience horizontal compression and the resultant vertical expansion. If we rescale the two models so that they have the same maximum slip, the model of full trench-breaking rupture will have a greater seismic moment but still a much lower vertical CSD (Wang and He 2007). This result demonstrates the importance of the

frictional behavior of the shallowest fault segment in affecting seafloor uplift. However, it addresses only one aspect of tsunamigenic CSD. Many other factors contribute to tsunami generation. For example, although the full trench-breaking model yields very low vertical CSD, its horizontal CSD may contribute to tsunami generation. If the seafloor slopes at angle α , its horizontal motion D in the slope direction should raise the seafloor by $D \tan \alpha$ relative to a fixed water column above, an effect addressed by Tanioka and Satake (1996). The speed of the coseismic slip is also an important factor in tsunami generation. In some rare cases, the rupture is too low to generate much seismic wave energy yet fast enough to generate rather large tsunamis, giving rise to a class of earthquake called tsunami earthquakes (Kanamori 1972).

Elastic deformation of the ocean floor as discussed above is the primary cause of tsunami generation during subduction earthquakes, but inelastic deformation can be locally important. For example, although the lower continental slope of active margins is on average the expression of a critically tapered Coulomb wedge, seafloor topography at these margins is rugged at smaller scales due to sedimentation, erosion, and deformation processes, and where the local surface slope is sufficiently high earthquake shaking may trigger gravitational failure. Such submarine “landslides” may have a locally significant effect on tsunami generation. Another potentially important inelastic process is the coseismic activation of out-of-sequence thrust faults (splay faults) in the accretionary prism (Fig. 1a). Splay faults are much more steeply dipping, and their thrust motion will serve to “redirect” the low-angle slip of the megathrust to a higher angle and thus may greatly enhance local seafloor uplift and contribute to tsunami generation (Fukao 1979; Park et al. 2002). As mentioned in the “Introduction”, from the continuum perspective, such faulting is a manifestation of Coulomb plasticity. At the local scale, it is actually frictional sliding of a contact surface with elastically deforming rocks on both sides. By comparison of the splay faults schematically illustrated in Fig. 1a and the potential failure planes of the

critical wedge in Fig. 2b, we can see that some of the splay faults are oriented optimally for thrust failure if the frontal wedge is compressed during a megathrust earthquake.

Future Directions

The connection between wedge mechanics and great earthquakes and tsunamis at subduction zones is an emerging new field of study. It leads to challenges in both theoretical development and experimental design and thus excellent research opportunities. We need better constraints on how stresses along different downdip segments of the subduction fault evolve with time throughout an earthquake cycle and how the evolution impacts wedge and seafloor deformation. A number of outstanding questions are to be addressed: Can we constrain the updip limit of the seismogenic zone using wedge morphology? What is the time-scale of stress relaxation along the updip segment of the megathrust after an earthquake? Does the seismogenic zone stay locked in the interseismic period? How does pore fluid pressure evolve in an earthquake cycle? How is the transfer of material from the incoming plate to the upper plate (accretion), from the subducted plate to the upper plate (underplating), or from the upper plate to the subducted plate (tectonic erosion) accomplished? What determines the dominant mode of material transfer? What does the spatial change in wedge morphology tell us about changes in the mechanical state of the wedge and the megathrust fault? These questions should be put in the proper context of larger-scale processes such as the viscoelastic relaxation of the mantle following a megathrust earthquake and the deformation of the subducting plate in earthquake cycles (Wang 2007).

Sandbox experiments designed to study wedge mechanics and dynamic friction experiments designed to study fault mechanics are traditionally separate research activities addressing processes of vastly different timescales. The linkage between subduction earthquakes and submarine wedge evolution suggests the need to combine these experiments. Rapid motion used to simulate

earthquakes has begun to be introduced into sand-box experiments (Rosenau et al. 2006).

The most promising type of field observation is continuous monitoring of deformation, such as strain and tilt, and fluid pressure using submarine borehole and seafloor observatories. Seafloor elevation change in response to the 2003 Tokachi-oki, northeast Japan, earthquake (M8.2), continuously recorded by two seafloor pressure sensors, clearly indicated coseismic strengthening behavior of the shallowest segment of the subduction fault (Baba et al. 2006). Formation fluid pressure changes detected at subsea borehole observatories at the Nankai Trough subduction zone, southwest Japan, have been interpreted to indicate transient aseismic motion of a part of the locked seismogenic zone and/or dynamics of the incoming plate (Davis et al. 2006). A number of very-low-frequency earthquakes have been remotely detected within the Nankai Trough accretionary prism using land-based seismic networks (Ito and Obara 2006), revealing the need for near-field observation using seafloor systems. Submarine monitoring in conjunction with land-based monitoring at subduction zones that are currently in different phases of earthquake cycles will allow us to understand the evolution of fault and wedge stresses during the interseismic period. In this regard, cabled seafloor monitoring networks including borehole observatories, being designed or implemented at different subduction zones (Juniper et al. 2006; Kaneda 2006) will surely yield valuable data in the near future.

Acknowledgments We thank EE Davis, N Kukowski, SE Lallemant, and K Sa-take for reviewing the article and providing valuable comments. This work is Geological Survey of Canada contribution 20070221.

Bibliography

Primary Literature

- Baba T, Hirata K, Hori T, Sakaguchi H (2006) Offshore geodetic data conducive to the estimation of the after-slip distribution following the 2003 Tokachi-oki earthquake. *Earth Planet Sci Lett* 241:281–292
- Barr TD, Dahlen FA (1990) Constraints on friction and stress in the Taiwan fold-and-thrust belt from heat flow and geochronology. *Geology* 18:111–115
- Breen NA, Orange DL (1992) The effects of fluid escape on accretionary wedges 1. Variable porosity and wedge convexity. *J Geophys Res* 97:9265–9275
- Briggs RW, Sieh K, Meltzner AJ, Natawidjaja D, Galetzka J, Suwargadi B, Hsu Y-J, Simons M, Hananto N, Suprihanto I, Prayudi D, Avoué J-P, Prawirodirdjo L, Bock Y (2006) Deformation and slip along the Sunda megathrust in the great 2005 Nias-Simeulue earthquake. *Science* 311:1897–1901
- Brown K, Kopf A, Underwood MB, Weinberger JL (2003) Compositional and fluid pressure controls on the state of stress on the Nankai subduction thrust: a weak plate boundary. *Earth Planet Sci Lett* 241:589–603
- Byerlee JD (1978) Friction of rocks. *Pure Appl Geophys* 116:615–626
- Byrne DE, Davis DM, Sykes LR (1988) Local and maximum size of thrust earthquakes and the mechanics of the shallow region of subduction zones. *Tectonics* 7:833–857
- Chapple WM (1978) Mechanics of thin-skinned fold-and-thrust belts. *Geol Soc Am Bull* 89:1189–1198
- Dahlen FA (1984) Noncohesive critical Coulomb wedges: an exact solution. *J Geophys Res* 89:10125–10133
- Dahlen FA, Suppe J, Davis DM (1984) Mechanics of fold-and-thrust belts and accretionary wedges: cohesive Coulomb theory. *J Geophys Res* 89:10087–10101
- Davis DM (1990) Accretionary mechanics with properties that vary in space and time. In: Debout GE et al (eds) *Subduction: top to bottom*. AGU Monograph 96, Washington, DC, pp 39–48
- Davis DM, Suppe J, Dahlen FA (1983) Mechanics of fold-and-thrust belts and accretionary wedges. *J Geophys Res* 88:1153–1172
- Davis EE, Becker K, Wang K, Obara K, Ito Y (2006) A discrete episode of seismic and aseismic deformation of the Nankai subduction zone accretionary prism and incoming Philippine Sea plate. *Earth Planet Sci Lett* 242:73–84
- Dewhurst DN, Clennell MB, Brown KM, Westbrook GK (1996) Fabric and hydraulic conductivity of sheared clays. *Géotechnique* 46:761–768
- Dieterich JH (1979) Modeling of rock friction: 1. Experimental results and constitutive equations. *J Geophys Res* 84:2161–2168
- Elliot D (1976) The motion of thrust sheets. *J Geophys Res* 81:949–963
- Enlow RL, Koons PO (1998) Critical wedges in three dimensions: analytical expressions from Mohr–Coulomb constrained perturbation analysis. *J Geophys Res* 103:4897–4914
- Fletcher RC (1989) Approximate analytical solutions for a cohesive fold-and-thrust wedge: some results for lateral variation in wedge properties and for finite wedge angle. *J Geophys Res* 94:10347–10354
- Fukao Y (1979) Tsunami earthquakes and subduction processes near deep-sea trenches. *J Geophys Res* 84:2303–2314

- Fuller CW, Willett SD, Brandon MT (2006) Formation of forearc basins and their influence on subduction zone earthquakes. *Geology* 34:65–68
- Geist EL, Dmowska R (1999) Local tsunamis and distributed slip at the source. *Pure Appl Geophys* 154: 485–512
- Hsu Y-J, Simons M, Avouac J-P, Galetzka J, Sieh K, Chlieh M, Natawidjaja D, Prawirodirdjo L, Bock Y (2006) Frictional afterslip following the 2005 Nias-Simeulue earthquake, Sumatra. *Science* 312: 1921–1926
- Hu Y, Wang K (2006) Bending-like behavior of wedge-shaped thin elastic fault blocks. *J Geophys Res*:111. <https://doi.org/10.1029/2005JB003987>
- Hyndman RD, Wang K (1993) Thermal constraints on the zone of major thrust earthquake failure: the Cascadia subduction zone. *J Geophys Res* 98:2039–2060
- Ito Y, Obara K (2006) Dynamic deformation of the accretionary prism excites very low frequency earthquakes. *Geophys Res Lett* 33. <https://doi.org/10.1029/2005GL025270>
- Juniper K, Bornhold B, Barnes C (2006) NEPTUNE Canada community science experiments. *EOS Trans Am Geophys Union* 87(52). Fall Meeting Supplement: Abstract OS34F-04
- Kanamori H (1972) Mechanism of tsunami earthquakes. *Phys Earth Planet Interior* 6:246–259
- Kaneda Y (2006) The advanced dense ocean floor observatory network system for mega-thrust earthquakes and tsunamis in the Nankai Trough – precise real-time observatory and simulating phenomena of earthquakes and tsunamis. *EOS Trans Am Geophys Union* 87(52). Fall Meeting Supplement: Abstract OS34F-01
- Kukowski N, von Hune R, Malavieille J, Lallemand SE (1994) Sediment accretion against a buttress beneath the Peruvian continental margin at 12° S as simulated with sandbox modeling. *Geol Rundsch* 83:822–831
- Lallemand SE, Schnürle P, Malavieille J (1994) Coulomb theory applied to accretionary and nonaccretionary wedges: possible causes for tectonic erosion and/or frontal accretion. *J Geophys Res* 99:12033–12055
- Lohrmann J, Kukowski N, Adam J, Oncken O (2003) The impact of analogue material properties on the geometry, kinematics, and dynamics of convergent sand wedges. *J Struct Geol* 25:1691–1711
- Moore DE, Lockner DA (2007) Friction of the smectite clay montmorillonite: a review and interpretation of data. In: Dixon T, Moore JC (eds) *The seismogenic zone of subduction thrust faults*. Columbia University Press, New York
- Moore JC, Saffer D (2001) Updip limit of the seismogenic zone beneath the accretionary prism of southwest Japan: an effect of diagenetic to low grade metamorphic processes and increasing effective stress. *Geology* 29:183–186
- Morone C (1998) Laboratory-derived friction laws and their application to seismic faulting. *Annu Rev Earth Planet Sci* 26:649–696
- Mourgues R, Cobbold PR (2006) Thrust wedges and fluid overpressures: sandbox models involving pore fluids. *J Geophys Research* 111. <https://doi.org/10.1029/2004JB003441>
- Park JO, Tsuru T, Kodaira S, Cummins PR, Kaneda Y (2002) Splay fault branching along the Nankai subduction zone. *Science* 297:1157–1160
- Rice J (2006) Heating and weakening of faults during earthquake slip. *J Geophys Res* 111. <https://doi.org/10.1029/2005JB004006>
- Rosenau M, Melnick D, Brookhagen B, Echtle HP, Oncken O, Strecker MR (2006) About the relationship between forearc anatomy and megathrust earthquakes. *EOS Trans Am Geophys Union* 87(52). Fall Meeting Supplement: Abstract T12C-04
- Ruina A (1983) Slip instability and state variable friction laws. *J Geophys Res* 88:10359–10370
- Sibson RH (2003) Thickness of the seismic slip zone. *Bull Seismol Soc Am* 93:1169–1178
- Smit JHW, Brun JP, Sokoutis D (2003) Deformation of brittle-ductile thrust wedges in experiments and nature. *J Geophys Res* 108. <https://doi.org/10.1029/2002JB002190>
- Takahashi M, Mizoguchi K, Kitamura K, Masuda K (2007) Effects of clay content on the frictional strength and fluid transport property of faults. *J Geophys Res* 112. <https://doi.org/10.1029/2006JB004678>
- Tanioka Y, Satake K (1996) Tsunami generation by horizontal displacement of ocean bottom. *Geophys Res Lett* 23:861–864
- von Huene R, Ranero CR (2003) Subduction erosion and basal friction along the sediment-starved convergent margin off Antofagasta, Chile. *J Geophys Res* 108. <https://doi.org/10.1029/2001JB001569>
- Vrolijk P (1990) On the mechanical role of smectite in subduction zones. *Geology* 18:703–707
- Wang CY (1980) Sediment subduction and frictional sliding in a subduction zone. *Geology* 8:530–533
- Wang K (2007) Elastic and viscoelastic models of subduction earthquake cycles. In: Dixon T, Moore JC (eds) *The seismogenic zone of subduction thrust faults*. Columbia University Press, New York
- Wang WH, Davis DM (1996) Sandbox model simulation of forearc evolution and noncritical wedges. *J Geophys Res* 101:11329–11339
- Wang K, He J (1999) Mechanics of low-stress forearcs: Nankai and Cascadia. *J Geophys Res* 104: 15191–15205
- Wang K, He J (2007) Effects of Frictional Behaviour and Geometry of Subduction Fault on Coseismic Seafloor Deformation. *Bull Seismol Soc Am* 98:571–579
- Wang K, Hu Y (2006) Accretionary prisms in subduction earthquake cycles: the theory of dynamic Coulomb wedge. *J Geophys Res* 111. <https://doi.org/10.1029/2005JB004094>
- Wang K, He J, Hu Y (2006) A note on pore fluid pressure ratios in the Coulomb wedge theory. *Geophys Res Lett* 33. <https://doi.org/10.1029/2006GL027233>

- Willett S, Beaumont C, Fullsack P (1993) Mechanical model for the tectonics of doubly vergent compressional orogens. *Geology* 21:371–374
- Xiao HB, Dahlen FA, Suppe J (1991) Mechanics of extensional wedges. *J Geophys Res* 96: 10301–10318
- Zhao WL, Davis DM, Dahlen FA, Suppe J (1986) Origin of convex accretionary wedges: evidence from Barbados. *J Geophys Res* 91:10246–10258

Books and Reviews

- Dahlen FA (1990) Critical taper model of fold-and-thrust belts and accretionary wedges. *Annu Rev Earth Planet Sci* 18:55–99
- Dixon T, Moore JC (eds) (2007) *The seismogenic zone of subduction thrust faults*. Columbia University Press, New York
- Scholz CH (2003) *The mechanics of earthquakes and faulting*, 2nd edn. Cambridge University Press, Cambridge, 471 p

Tsunamis, Inverse Problem of

Kenji Satake

Earthquake Research Institute, The University of Tokyo, Tokyo, Japan

Article Outline

Glossary

Definition of the Subject

Introduction

Tsunami Generation by Earthquakes

Tsunami Propagation

Numerical Computations

Tsunami Observations

Estimation of Tsunami Source

Estimation of Earthquake Fault Parameters

Future Directions

Bibliography

Glossary

Fault parameters Earthquake source is modeled as a fault motion, which can be described by nine static parameters. Once these fault parameters are specified, the seafloor deformation due to faulting, or initial condition of tsunamis, can be calculated by using elastic dislocation theory.

Inverse problem Unlike a forward problem, which starts from a tsunami source then computes propagation in the ocean and predicts travel times and/or water heights on coasts, an inverse problem starts from tsunami observations to study the generation process. While forward modeling is useful for tsunami warning or hazard assessments, inverse modeling is a typical approach for geophysical problems.

Refraction and inverse refraction diagrams (travel-time map) Refraction diagram is a

map showing isochrons or lines of equal tsunami travel times calculated from the source toward coasts. Inverse refraction diagram is a map showing arcs calculated backward from observation points. The tsunami source can be estimated from the arcs corresponding to tsunami travel times.

Shallow water (long) waves In hydrodynamics, water waves can be treated as shallow water or long waves when the wavelength is much larger than the water depth. In such a case, the entire water mass from water bottom to surface moves horizontally, and the wave propagation speed is given as a square root of the product of the gravitational acceleration and the water depth.

The 2004 Indian Ocean tsunami On December 26, 2004, a gigantic earthquake, the largest in the last half century in the world, occurred off the west coast of Sumatra Island, Indonesia. With the source extending more than 1000 km through Nicobar and Andaman Islands, the earthquake generated tsunami which attacked the coasts of Indian Ocean and caused the worst tsunami disaster in history. The total casualties were about 230,000 affecting countries as far away as Africa.

The 2011 Tohoku, Japan, tsunami On March 11, 2011, another gigantic earthquake, the largest in Japan's history, occurred off Tohoku, Japan. This earthquake generated devastating tsunami with a maximum coastal height of nearly 40 m and flooding distance of nearly 5 km from the coast. The earthquake and tsunami caused nearly 20,000 casualties and serious damage to Fukushima nuclear power station.

Definition of the Subject

Forward modeling of tsunami starts from given initial condition, computes its propagation in the

ocean, and calculates tsunami arrival times and/or water heights on coasts. Once the initial condition is provided, the propagation and coastal behavior can be numerically computed on actual bathymetry (Fig. 1).

Recent technological developments make it possible to carry out tsunami forward modeling with speed and accuracy usable for the early tsunami warning or detailed hazard assessments. However, the initial condition, or the tsunami generation process, is still poorly known, because large tsunamis are rare and the tsunami generation in the ocean is not directly observable. Indirect estimation of tsunami source, mostly on the basis of seismological analyses, is used as the initial condition of tsunami forward modeling. More direct estimation of tsunami source is essential to better understand the tsunami generation process and to more accurately forecast the tsunami on coasts.

Inverse modeling of tsunami starts from observed tsunami data, to study the tsunami source. The propagation process can be evaluated by using numerical simulation, as in the forward modeling. As the observed tsunami data, tsunami arrival times, heights, or waveforms recorded on instruments are used. For historical tsunamis, tsunami heights can be estimated from description of damage on historical documents. For prehistoric tsunamis, geological studies of tsunami deposits

can be used to estimate the coastal tsunami heights or flooding areas.

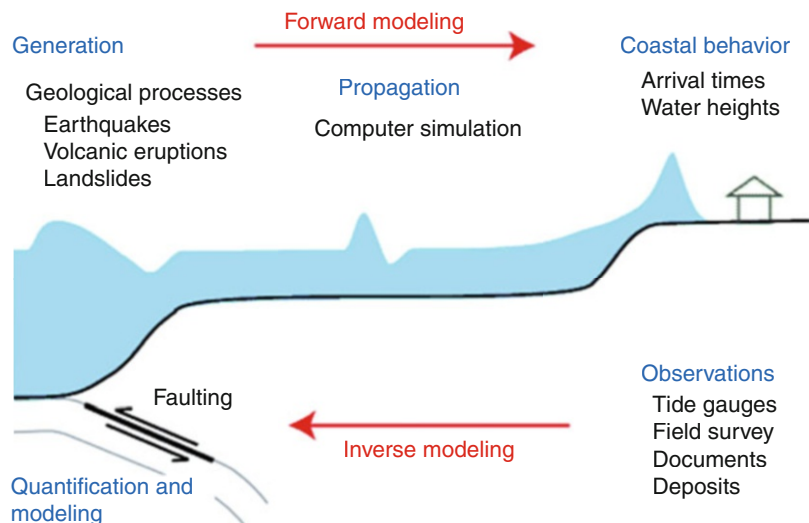
Introduction

Tsunamis are oceanic gravity waves generated by seafloor deformation due to submarine earthquakes or other submarine geological processes such as volcanic eruptions, landslides, or asteroid impacts. While earthquake tsunamis, such as the 2004 Indian Ocean tsunami caused by the Sumatra-Andaman earthquake or the 2011 Japan tsunami caused by the Tohoku earthquake, are most frequent, large volcanic eruptions such as the 1883 Krakatau eruption off Sumatra Island also cause ocean-wide tsunamis. Landslides, often triggered by earthquakes, cause locally large tsunamis, but the effects are usually limited to the area around the source.

Most tsunamigenic geological processes produce seafloor deformation. When horizontal scale, or wavelength, of the seafloor deformation is much larger than the water depth, a similar disturbance appears on the water surface and becomes the source of tsunami. This is called shallow water, or long-wave, approximation. For large earthquakes, wavelength of seafloor deformation is an order of several tens to hundreds of km, while the ocean depth is up to several km;

Tsunamis, Inverse Problem of,

Fig. 1 Schematic diagram showing tsunami generation, propagation, and coastal behavior. Forward modeling starts from tsunami source and forecasts the coastal behavior, while inverse modeling starts from observed data to estimate the tsunami source



hence, the long-wave approximation is valid. For small-scale disturbance relative to water depth, such as submarine landslides or volcanic eruptions in deep seas, the shallow water approximation may not be valid.

This entry reviews inverse methods to study tsunami sources from the observations, with the 2004 Indian Ocean tsunami and the two Japanese tsunamis caused by the 1968 Tokachi-oki and 2011 Tohoku earthquakes as examples. Section “[Tsunami Generation by Earthquakes](#)” describes the tsunami generation by earthquakes, with emphasis on the fault parameters and their effects on tsunamis. Section “[Tsunami Propagation](#)” describes tsunami propagation: shallow water theory and numerical computation. Section “[Tsunami Observations](#)” summarizes the tsunami observation: instrumental sea-level data and run-up height estimates for modern, historical, and prehistoric tsunamis. Section “[Estimation of Tsunami Source](#)” describes methods of modeling and quantifying tsunami source and of analyzing tsunami travel times, amplitudes, and waveforms, including some historical developments. Section “[Estimation of Earthquake Fault Parameters](#)” focuses on the estimation of earthquake fault parameters, including the waveform inversion of tsunami data to estimate heterogeneous fault motion and its application for tsunami warning.

Tsunami Generation by Earthquakes

Fault Parameters and Seafloor Deformation

The seafloor deformation due to earthquake faulting can be calculated by using the elastic

theory of dislocation. The displacement, u_k , in an infinite homogeneous medium due to dislocation Δu_i across surface Σ is given by the Volterra’s theorem as (Steketee 1958)

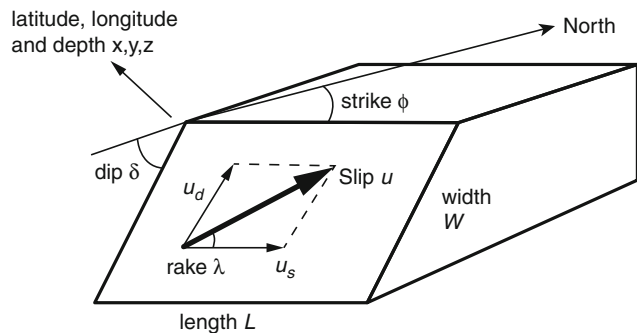
$$u_k = \frac{1}{8\pi\mu} \int_{\Sigma} \Delta u_i \left\{ \lambda \delta_{ij} u_k^{,n} + \mu (u_k^{,j} + u_j^{,k}) \right\} v_j dS \quad (1)$$

where λ and μ are Lamé constants, δ_{ij} is Kronecker’s delta, and v is the unit normal to the surface. The expression u_i^j denotes the i th component of the displacement due to the j th component of a point force at the source whose magnitude is $8\pi\mu$, and $u_i^{j,k}$ indicates its spatial derivative with respect to the k th coordinate. For a half-space with free surface, a mirror image can be used to cancel the stress components on the free surface. The explicit formulas are given by Mansinha and Smylie (1971) or Okada (1985).

The fault parameters needed to compute surface deformation are summarized in Fig. 2. They are fault length (L), width (W), strike (ϕ), dip (δ), rake (λ), slip amount (u), and location (x, y, z). The slip u can be decomposed into strike-slip component u_s and dip-slip component u_d . The strike ϕ is measured clockwise from north, dip angle δ is downward from horizontal, and rake angle λ is a movement of hanging wall measured counter-clockwise from horizontal (see Fig. 2). Therefore, the fault motion is reverse if $\lambda > 0^\circ$ and normal if $\lambda < 0^\circ$. The fault motion has left-lateral component if $|\lambda| < 90^\circ$ and right-lateral component if $|\lambda| > 90^\circ$.

The physical parameter to quantify the fault motion is the seismic moment, M_0 , defined as

Tsunamis, Inverse Problem of, Fig. 2 Fault parameters. Seafloor deformation can be computed from these static parameters



$$M_0 = \mu u S = \mu u L W \quad (2)$$

where μ is rigidity around the fault. More conventional parameter of earthquake size is a magnitude scale, which has been determined from amplitudes of seismograms. To relate the seismic moment and magnitude scales, the moment magnitude scale, M_w , is defined as (Hanks and Kanamori 1979; Kanamori 1977)

$$M_w = \frac{2}{3} \log M_0 - 10.7 \quad (3)$$

where M_0 is given in dyne.cm (10^{-7} Nm).

Most of the above fault parameters can be estimated from seismic wave analysis. The location and depth of fault (x, y, z) correspond to hypocenter, which is estimated from arrival times of seismic waves. The fault geometry (ϕ, δ, λ) is estimated from the polarity distribution of body wave first motions or azimuthal distribution of surface wave amplitudes. The seismic moment is estimated from waveform modeling of seismic waves. The fault sizes L and W are more difficult to estimate; they are usually estimated from after-shock distribution or detailed waveform modeling of seismic body waves. The slip amount, u , is indirectly estimated, from seismic moment M_0 , by assuming μ and fault size (L and W). All such estimates assume that the faulting is planar and continuous, which most often is a simplification of real, more complex faulting. Besides the above direct estimations, the scaling relations among the fault parameters obtained from many past earthquakes can be used to infer some parameters from the others (Murotani et al. 2013).

The 2004 Sumatra-Andaman earthquake was the largest earthquake since the 1960 Chilean earthquake (M_w 9.5) or 1964 Alaskan earthquake (M_w 9.2). The seismic moment estimates range $4 - 12 \times 10^{22}$ Nm, and the corresponding moment magnitude M_w ranges 9.0–9.3 from the seismological analyses (Lay et al. 2005; Stein and Okal 2005; Tsai et al. 2005). The aftershock area extended from off Sumatra through the Nicobar to Andaman Islands with the total fault length of 1200–1300 km (Lay et al. 2005). The seismic moment of the 2011 Tohoku earthquake is

estimated as $\sim 4 \times 10^{22}$ Nm from various analyses (seismic waves, land and marine geodetic data, and tsunami waveforms). The corresponding M_w is 9.0. For such gigantic earthquakes, multiple fault planes with different strike and slip amounts are needed to represent the fault motion, as shown later (section “[Estimation of Earthquake Fault Parameters](#)”).

Effect of Fault Parameters on Tsunami Generation

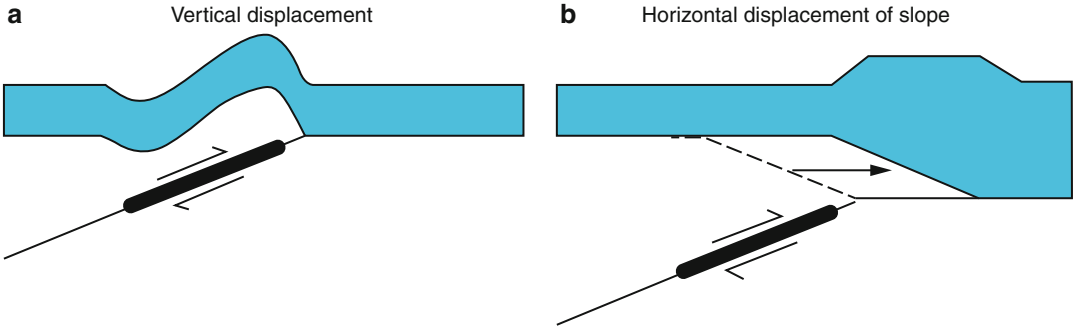
Among the above static fault parameters, the slip amount has the largest effect on the vertical seafloor deformation and the tsunami amplitude. The dip angle and fault depth are also important parameters to control tsunami amplitude (Geist 1998; Yamashita and Sato 1974). Dynamic parameters such as rupture velocity are found to be insignificant for tsunami generation. However, for a gigantic earthquake such as the 2004 Sumatra-Andaman earthquake with the source length over 1000 km, rupture propagation effect on tsunamis is not negligible (Fujii and Satake 2007).

Amplitude of far-field seismic waves, either body waves or surface waves, is controlled by seismic moment, while amplitude of tsunami is controlled by fault slip. Satake and Tanioka (2003) found for the 1998 Papua New Guinea tsunami that the far-field tsunami amplitudes are proportional to the volume of displaced water, while the near-field tsunami amplitudes are controlled by the potential energy of the displacement.

Traditionally, only the vertical component of seafloor deformation has been considered for tsunami generation (Fig. 3a). If an earthquake occurs on a steep ocean slope such as trench slope, horizontal displacement due to faulting moves the slope and contributes to the tsunami generation (Tanioka and Satake 1996). The effective vertical movement (positive upward) due to faulting can be written as follows (Fig. 3b):

$$u_z + u_x \frac{\partial H}{\partial x} + u_y \frac{\partial H}{\partial y} \quad (4)$$

where u_z is vertical component and u_x and u_y are horizontal components of seafloor deformation



Tsunamis, Inverse Problem of, Fig. 3 Seafloor deformation and tsunami source. (a) Vertical seafloor deformation becomes the tsunami source. (b) When the seafloor is

not flat, horizontal displacement of the slope also affects the tsunami generation (Tanioka and Satake 1996)

and H is water depth (measured positive downward). For the 2011 Tohoku tsunami, the effect of horizontal displacement is as large as 20–40% of the observed tsunami amplitudes (Satake et al. 2013).

Tsunami Propagation

Shallow Water Theory

The equation of motion, or conservation of momentum, for shallow water, or long-wave, theory is given as follows:

$$\frac{\partial \mathbf{V}}{\partial t} + (\mathbf{V} \cdot \nabla) \mathbf{V} = -g \nabla h + C_f \frac{\mathbf{V} |\mathbf{V}|}{d + h} \quad (5)$$

where $\mathbf{V}$ is the depth-averaged horizontal velocity vector, h is the water height or tsunami amplitude, d is the water depth, and g is the gravitational acceleration. On the left-hand side, the first term represents local acceleration, and the second term represents nonlinear advection. On the right-hand side, the first term represents pressure gradient, or restoring force due to gravity, and the second term represents nonlinear bottom friction where C_f is the nondimensional frictional coefficient.

The equation of continuity, or conservation of mass, can be written as

$$\frac{\partial (d + h)}{\partial t} = -\nabla \cdot \{(d + h) \mathbf{V}\}. \quad (6)$$

These equations can be linearized as

$$\frac{\partial \mathbf{V}}{\partial t} = -g \nabla h \quad (7)$$

$$\frac{\partial h}{\partial t} = -\nabla \cdot (d \mathbf{V}) \quad (8)$$

when the tsunami amplitude h is small compared to water depth d and the bottom friction can be neglected. Such an assumption is valid for deep ocean or most of the tsunami propagation path. Near the coasts, nonlinear terms play important roles; hence, linearization may not be valid. The major advantage of the linear theory is the superposition principle; the computational results can be easily scaled to estimate with different initial water heights.

From Eqs. 7 and 8, the wave equation with wave velocity $\sqrt{gd}$ can be derived. This indicates that the tsunami speed is controlled by water depth. Once the water depth distribution, or ocean bottom bathymetry, is known, then the tsunami propagation can be computed numerically.

Numerical Computations

Equations 5 and 6, or 7 and 8 for linearized case, can be directly solved by numerical methods, once the initial condition is given. The tsunami wave velocity distribution, which is given by the bathymetry, is much better known than the velocity distribution of seismic waves; hence, actual values can be used. Finite-difference method with staggered grids is popularly used (Mader 1988;

Intergovernmental Oceanographic Commission 1997; Imamura 2009), while the use of other methods such as finite-element methods has been also proposed (Yeh et al. 1996). Grid size for finite-difference computations is typically a few km for deep ocean, but grids as fine as tens to hundreds of meters are used near coasts. The temporal changes in water height at grid points corresponding to the observation points are used as computed tsunami waveforms.

The database of global water depth or bathymetry data such as ETOPO1 (NOAA/NGDC) or GEBCO (British Oceanographic Data Centre) are popularly used. The ETOPO database is based on predicted bathymetry from satellite altimetry data (Smith et al. 2005) with interval of 1 min (about 1.8 km), while GEBCO data are digitized from nautical charts with grid interval of 15 arc seconds. Higher-resolution bathymetry data near coasts are open to public in some countries such as the USA (NOAA/NGDC) or Japan (JODC).

Tsunami Observations

Instrumental Data

Traditional instrumental data for tsunami observation are tide gauge records. Tide gauges are typically installed on ports or harbors to define datum or to monitor ocean tides. The temporal resolution is usually low with a sampling interval of several minutes or longer. For the tsunami monitoring, higher sampling rate, at least 1 min or shorter interval, is required. While the recorded tsunami waveforms contain coastal effects such as coastal reflections or resonance particularly for the later phase, the initial part of tsunami signals is more dominant by the tsunami source effect; hence, the source information can be retrieved. Currently, sea-level measurement data from many tide gauge stations are transmitted through weather satellite and available in real time. Figure 4 (left) shows some of tide gauge records of the 2004 Indian Ocean tsunami.

Tsunami waveforms are simpler offshore or in deep ocean, free from nonlinear coastal effects, though the signal is smaller. Offshore and deep

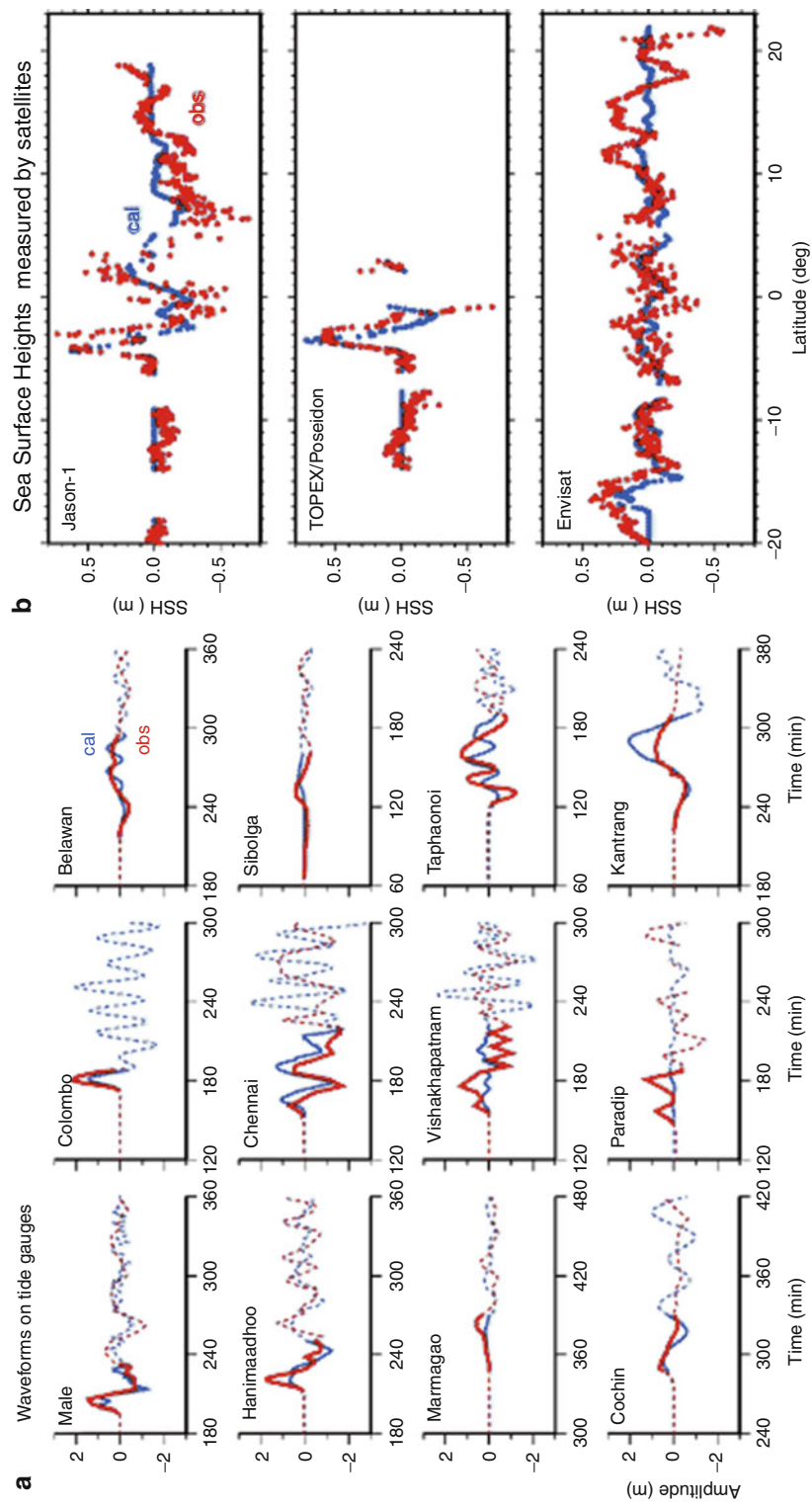
ocean tsunami observation facilities have been significantly developed recently. Offshore tsunami gauges such as GPS tsunami gauge (Kato et al. 2000) or cabled bottom pressure gauges (Mikada et al. 2006) have been developed and deployed around Japan, and they recorded tsunami waveforms. For the recent 2011 Tohoku tsunami in Japan, in addition to the GPS gauges, several cabled bottom pressure gauges have recorded the tsunami generation process around the source area (Fujii et al. 2011). The US NOAA deployed bottom pressure gauges, called Deep-ocean Assessment and Reporting of Tsunamis (DART) or simply tsunameters (Gonzalez et al. 2005). The bottom pressure signals are sent to surface buoys via acoustic telemetry in the ocean and then to land station via satellite. As described later, the DART data are used for real-time data assimilation and tsunami warning purposes (Titov et al. 2005). After the 2004 Indian Ocean tsunami, several tens of DART-type bottom pressure gauges have been deployed in the Pacific and Indian oceans.

Satellite altimeters captured the propagation of the 2004 Indian Ocean tsunami (Fig. 4). Three satellites flew over the Indian Ocean at a few hours after the earthquake and measured the sea surface height (SSH) of about 0.8 m in the middle of the Indian Ocean. The tsunami amplitudes in deep ocean are much smaller than the maximum coastal heights of more than 10 m. The SSH data are used to study the tsunami source (Fujii and Satake 2007; Hirata et al. 2006).

Modern, Historical, and Prehistoric Tsunami Heights

After damaging tsunamis, tsunami height distribution is often measured by survey teams (Synolakis and Okal 2005). Measurements are usually made for flow depth above ground, on the basis of various watermarks, and then converted to inundation height above sea level (Intergovernmental Oceanographic Commission 1998). The tsunami inundation heights are usually not constant along a profile from beach, and the height at most inland point is called run-up height.

For historical tsunamis, coastal tsunami heights can be estimated from descriptions of



Tsunamis, Inverse Problem of, Fig. 4 The sea-level data from the 2004 Indian Ocean tsunami (Fuji and Satake 2007). (a) Tsunami waveforms on tide gauges. Red curves indicate observed waveforms, and blue ones are computed. Data shown in solid lines are used for the waveform inversion. (b) Sea surface heights measured by three different satellites (see Fig. 5 for the tracks). Red shows the observed data, and blue is for computed surface heights

tsunami or its damage recorded in historical documents. Such estimates include various assumptions on sea levels and a relationship between tsunami damage and flow depth, but provide important tsunami data for historical tsunamis. For example, date and size of the last gigantic earthquake in the Cascadia subduction zone off North America were estimated as January 26, 1700 and $M_W \sim 9.0$ from the Japanese tsunami data (Satake et al. 2003).

Geological traces such as tsunami deposits can also be used to estimate ages and tsunami-affected areas for prehistoric tsunamis. In the last few decades, many studies of tsunami deposits, combined with numerical computations, have been made to analyze prehistoric tsunamis (Atwater et al. 2005; Dawson and Shi 2000). For example, in Hokkaido, prehistoric tsunami deposits indicate past tsunamis with larger inundation area and longer recurrence interval than those from the recent plate-boundary earthquakes along the southern Kuril trench, which were attributed to multi-segment earthquakes with $M_W \sim 8.4$ (Nanayama et al. 2003). The 869 Jogan earthquake, a predecessor of the 2011 Tohoku earthquake, has been studied from tsunami deposits, and the fault parameters were estimated before the 2011 tsunami (Sawai et al. 2012). The 2011 Tohoku tsunami left numerous sand deposits in Sendai plain, which provided valuable data for modern analog of paleotsunami studies.

Estimation of Tsunami Source

Refraction Diagram

Tsunami propagation can be computed and described as a refraction diagram or travel-time diagram. When the tsunami wavelength is smaller than the scale length of velocity heterogeneity, or the water depth variation is smooth, then the geometrical ray theory of optics can be applied. The wave fronts of propagating tsunami can be drawn on the basis of Huygens' principle. Alternatively, propagation of rays, which is orthogonal to wave fronts, can be traced from an assumed source. While refraction diagram does not provide information on water height, the relative amplitudes

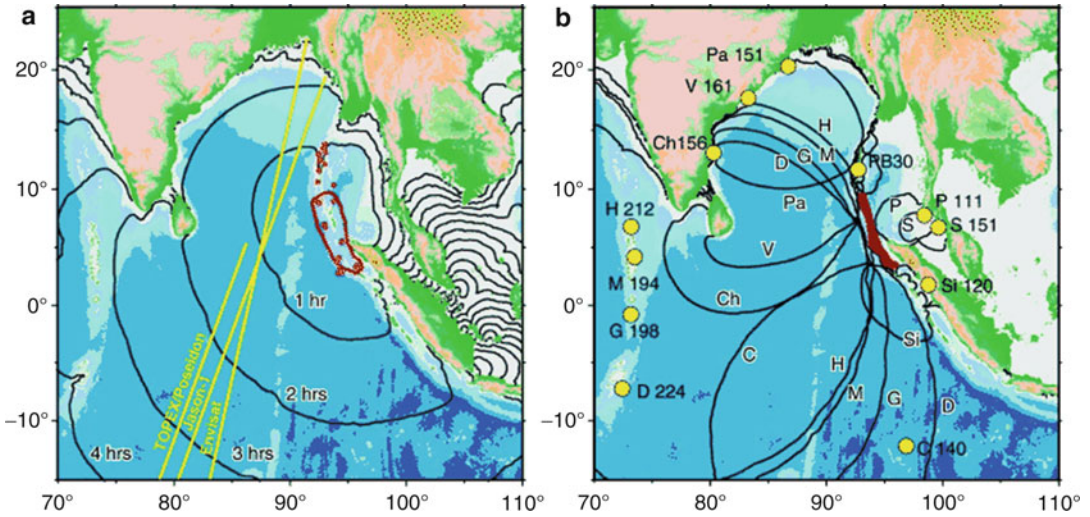
can be estimated from density of rays (Satake 1988).

Travel-time diagrams can be prepared for major tsunami sources and used for tsunami warning; as soon as the epicenter is known, the tsunami arrival times can be readily calculated. The refraction diagrams are usually drawn from a point source, but it is possible to draw it from an extended source for a great or giant earthquake. Figure 5a shows the refraction diagram from the 2004 Sumatra-Andaman earthquake with wave fronts at each hour. To the east of the assumed source, the tsunami is expected to arrive at the Thai coast in about 2 h through Andaman Sea. To the west, through deeper Bay of Bengal, the tsunami is expected to arrive at Sri Lanka also in 2 h. The predicted tsunami arrival times are similar to the actually observed values (Rabinovich and Thomson 2007).

Inverse Refraction Diagram

Refraction diagram can be drawn backward from coasts. Such a diagram is called inverse refraction diagram and is used to estimate the tsunami source area. When the tsunami travel time, that is, tsunami arrival time minus earthquake origin time, is known, the corresponding wave front, or travel-time arc, drawn from the tsunami observation point (typically tide gauge stations) would indicate the initial wave front at the tsunami source. The tsunami inverse refraction diagram was first drawn for the 1933 Sanriku tsunami (Miyabe 1934), although the estimated tsunami source was much larger than modern estimates, because both tsunami travel times and the bathymetry were poorly known.

The 2004 Indian Ocean tsunami was observed at many tide gauge stations in the Indian Ocean (Merrifield et al. 2005; Nagarajan et al. 2006). The tsunami arrival times were read from the tide gauge records, and tsunami travel times were calculated from the earthquake origin time. The tsunami propagation was then computed from each tide gauge station, and wave fronts corresponding to the travel time were drawn as travel-time arcs (Fig. 5b). These travel-time arcs surround the tsunami source, and the source area was estimated



Tsunamis, Inverse Problem of, Fig. 5 (a) Tsunami refraction diagram for the 2004 Sumatra-Andaman earthquake. Red dots indicate aftershocks within 1 day according to USGS. The red curve shows the assumed tsunami source. Tracks of three satellites with altimeters are shown by yellow lines. Black curves indicate tsunami wave fronts at each hour after the earthquake. (b) Tsunami

inverse refraction diagram for the same event. Station code and tsunami arrival times (in min) are attached to tide gauge stations (yellow circles) where the tsunami was instrumentally recorded. Black curves are the travel-time arcs computed for each station. Red area indicates inferred tsunami source

as about 900 km long (Lay et al. 2005; Neetu et al. 2005).

Estimation of Tsunami Source

Tsunami data can be used to study earthquake source processes in a similar way that seismic waves are used. This was first demonstrated for the 1968 Tokachi-oki earthquake ($M_0 = 2.8 \times 10^{21}$ Nm or $M_w = 8.3$) (Abe 1973). The tsunami source area estimated from an inverse refraction diagram agrees well with the aftershock area (Fig. 6a, b). In addition, the initial water surface disturbance was estimated as uplift at the south-eastern edge and subsidence at the northwestern edge, from the first motion of recorded tsunami waveforms on tide gauges. This pattern is very similar to the vertical bottom deformation due to the faulting, which was independently estimated from seismological analysis (Fig. 6c).

Green's Law and Tsunami Heights

The water height in the tsunami source area can be estimated from the observed tsunami heights along the coasts, by using the Green's law. The

Green's law is derived from the conservation of potential energy along rays (Mei 1989):

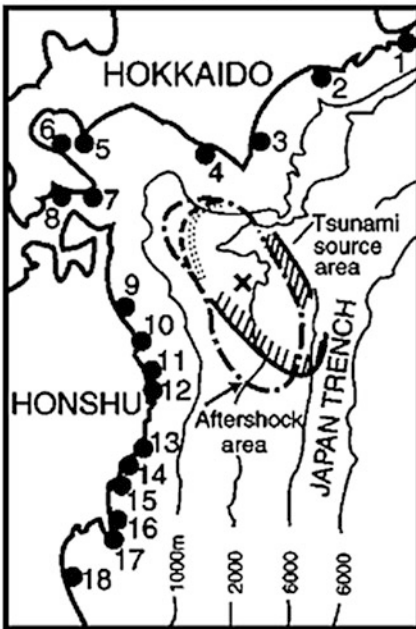
$$b_0 d_0^{1/2} h_0^2 = b_1 d_1^{1/2} h_1^2 \quad (9)$$

where d is the water depth, b is the distance between the neighboring rays, h is the tsunami amplitude, and the subscripts 0 and 1 indicate two different locations on the same ray. If the tsunami amplitude at location 0 (e.g., on the coast) is known, the tsunami amplitude at location 1 (e.g., at the source) can be estimated as

$$h_1 = \left(\frac{b_0}{b_1} \right)^{1/2} \left(\frac{d_0}{d_1} \right)^{1/4} h_0. \quad (10)$$

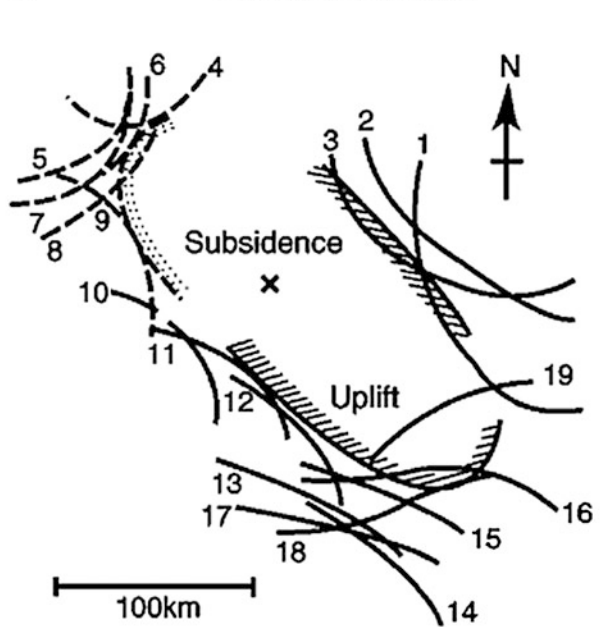
The ratio b_0/b_1 represents the spreading of rays, which can be graphically obtained from refraction diagrams. For the Tokachi-oki earthquake, the average tsunami height at the source was estimated as 1.8 m using the Green's law, which is very similar to 1.6 m, the average vertical seafloor displacement computed from the fault model (Abe 1973).

a 1968 Tokachi-oki earthquake

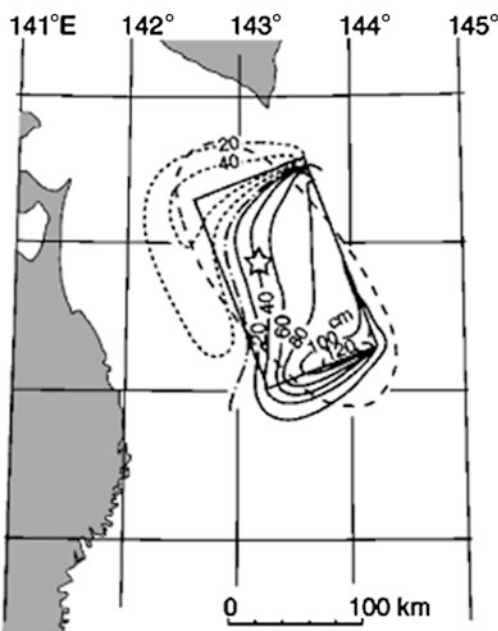


b

Tsunami source area

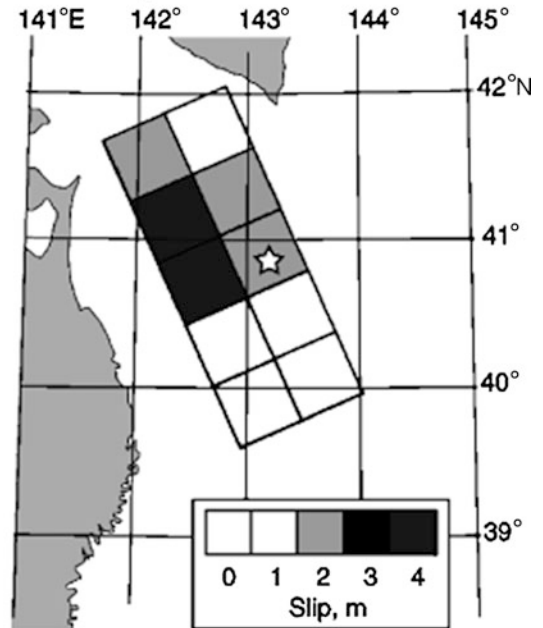


c Seafloor deformation



d

Slip distribution on fault



Tsunamis, Inverse Problem of, Fig. 6 The tsunami source area and seafloor deformation of the 1968 Tokachi-oki earthquake (Abe 1973; Satake 1989). (a) Estimated tsunami source, aftershock area, and distribution of tide gauge stations. (b) Travel-time arcs drawn by inverse refraction diagram. The numbers correspond to tide gauge

stations in (a). Solid and dashed curves show uplift and subsidence, respectively. (c) Seafloor deformation computed from a seismological fault model. (d) Slip distribution on fault estimated by an inversion of tsunami waveforms

The Green's law is also used to estimate the shoaling effects. For plane waves approaching the coast, the spreading ratio is unity; hence, the amplitude is proportional to a 1/4 power of water depth change. For example, when the water depth becomes a one hundredth, e.g., from 1000 to 10 m, the amplitude becomes three times larger.

Tsunami Magnitude

Tsunami magnitude scale, M_t , was introduced to quantify earthquake source that generated tsunamis (Abe 1979). The formulas were calibrated with the moment magnitude scale, M_w , of earthquakes. It is different from other tsunami magnitude or intensity scales that simply quantify the observed tsunamis. The definition of M_t for a trans-Pacific tsunami is (Abe 1979)

$$M_t = \log H + C + 9.1 \quad (11)$$

and for a regional ($100 \text{ km} < \Delta < 3500 \text{ km}$), tsunami is (Abe 1981)

$$M_t = \log H + \log \Delta + 5.8 \quad (12)$$

where H is a maximum amplitude on tide gauges in meters, C is a distance factor depending on a combination of the source and the observation points, and Δ is the nautical distance in km. The tsunami magnitude M_t was assigned as $M_t = 8.2$ for the 1968 Tokachi-oki earthquake, $M_t = 9.0$ for the 2004 Sumatra-Andaman earthquake, and $M_t = 9.1$ for the 2011 Tohoku earthquake.

Because the tsunami magnitude scale M_t is defined from tsunami amplitudes, it can be used to characterize "tsunami earthquakes" that produce much larger tsunamis than expected from seismic waves (see ► "Tsunami Earthquakes"). Abe (1989) defined "tsunami earthquakes" for such events with tsunami magnitude M_t larger than surface wave magnitude M_s by more than 0.5. It should not be confused with "tsunamigenic earthquake" which refers to any earthquake that generates tsunami.

Estimation of Earthquake Fault Parameters

For earthquake tsunamis, the fault parameters can be estimated by inverse modeling of tsunamis.

Such attempts were first made by a trial and error approach. In order to estimate the heterogeneous fault parameters, inversion of tsunami waveforms or run-up heights has been introduced.

Trial and Error Approach

Numerical simulation of tsunami has been carried out for many tsunamigenic earthquakes around Japan (Aida 1978). For the 1968 Tokachi-oki earthquake, tsunami waveforms were computed from two models, one based on seismological analysis (Fig. 6c) and another horizontally shifted by 28 km, and were compared with the observed tsunami waveforms recorded on tide gauges. Comparison of waveforms indicates that the latter model, shifted from that based on seismological analysis, shows better match in terms of tsunami arrival times. The slip amount on the fault was estimated as 4 m.

The best fault models are judged by comparison of the observed and computed tsunami waveforms. A few statistical parameters used to quantify the comparison are geometric mean, logarithmic standard deviation, and correlation coefficient. The geometric mean K of the amplitude ratio O_i/C_i , where O_i and C_i are the observed and computed tsunami amplitudes at station i , is given as

$$\log K = \frac{1}{n} \sum_i \log \frac{O_i}{C_i} \quad (13)$$

The logarithmic standard deviation κ is defined as

$$\log \kappa = \left[\frac{1}{n} \sum_i \left(\log \frac{O_i}{C_i} \right)^2 - (\log K)^2 \right]^{1/2} \quad (14)$$

If the logarithmic amplitude ratios obey the normal distribution $N(\log K, \log \kappa)$, then parameter κ can be considered as an error factor, because its logarithm shows the standard deviation. The geometric mean K indicates the relative size of the observed and computed tsunami models. The logarithmic standard deviation κ indicates the goodness of the model; the smaller κ means the better model. The arrival times of observed and computed waveforms are also compared, as indicated in the above example. Another parameter is correlation coefficient of the observed and computed

waveforms, which are also used for the comparison of models.

While the above parameters (K and κ) were originally defined for maximum amplitudes of waveforms, they are also used for comparison of observed and computed run-up heights. For tsunami hazard evaluation of nuclear power plants in Japan, tsunami source models need to satisfy $0.95 < K < 1.05$ and $\kappa < 1.45$ for the observed and computed coastal heights (Yanagisawa et al. 2007).

Heterogeneous Fault Motion

Seismological studies of large earthquakes have indicated that the slip amount on faults is not uniform but heterogeneous. For the 1968 Tokachi-oki earthquake, inversion of far-field body waves or regional Rayleigh waves showed the slip distributions similar to those estimated from tsunami waves (Kikuchi and Fukao 1985; Mori and Shimazaki 1985). The large slip area, sometimes called seismic asperity, produces high-frequency seismic waves thus important for strong-motion prediction for earthquake hazard assessments. The asperity produces large seafloor deformation; hence, it is also important for tsunami generation and its hazard estimation. Lay and Kanamori (1981) suggested that characteristic size of asperities differs from one subduction zone to another and is controlled by geological setting. Yamanaka and Kikuchi (2004), from studies of recurrent earthquakes off northern Honshu, showed that the same asperity ruptures in repeated earthquakes. Their asperity map can be used for earthquake and tsunami hazard assessment. Global studies of giant earthquakes indicate that the area and slip of asperities are roughly constant (Murotani et al. 2013).

Waveform Inversion

The asperity distribution can be estimated by an inversion of tsunami waveforms. In this method (Fig. 7), the fault plane is first divided into several subfaults, and the seafloor deformation is computed for each subfault with a unit amount of slip. Using these as the initial conditions, tsunami propagation is numerically computed for actual bathymetry, and the waveforms at tide gauge

stations, called Green's functions, are computed. Assuming that the tsunami generation and propagation are linear process, the observed tsunami waveforms are expressed as a linear superposition of Green's functions as follows:

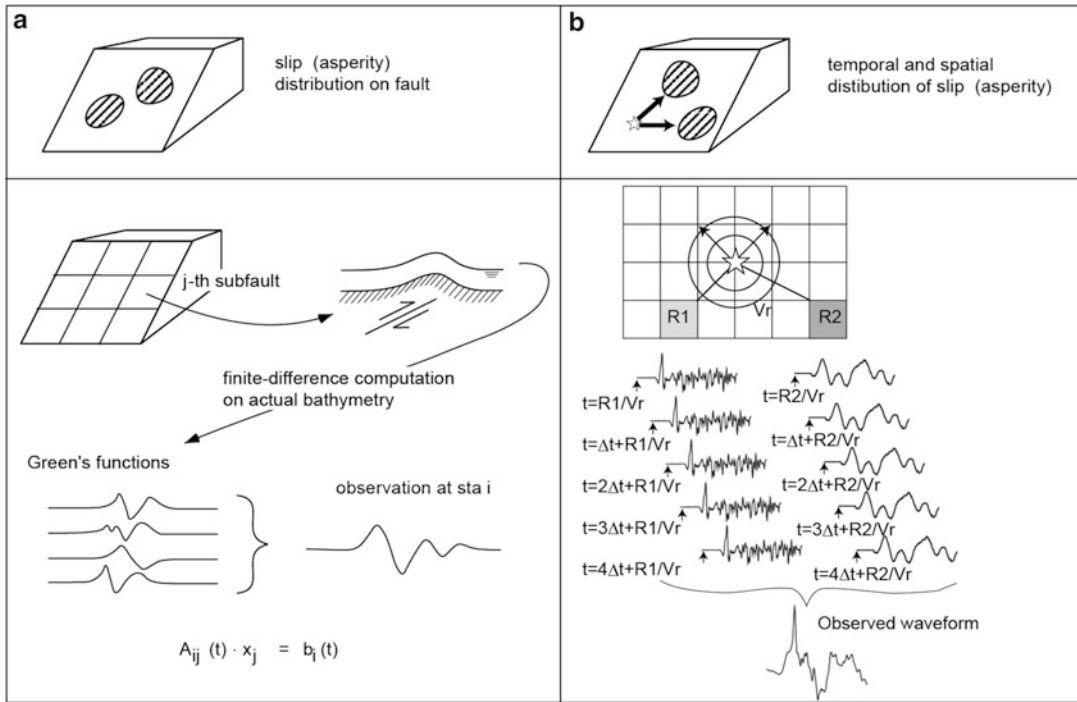
$$A_{ij}(t) \cdot x_j = b_i(t) \quad (15)$$

where $A_{ij}(t)$ is the computed waveform as a function of time t , or Green's function, at the i th station from the j th subfault; x_j is the amount of slip on the j th subfault; and $b_i(t)$ is the observed tsunami waveform at the i th station. The slip x_j on each subfault can be estimated by a least-square inversion of the above set of equations, by minimizing the $/_2$ norm of the residuals J :

$$J = \|\mathbf{A} \cdot \mathbf{x} - \mathbf{b}\| \rightarrow \min \quad (16)$$

where $\mathbf{A}$, $\mathbf{x}$, and $\mathbf{b}$ indicate matrix representations of elements in Eq. 15. Figure 6d shows the slip distribution on the fault for the 1968 Tokachi-oki earthquake. The source fault was divided into ten subfaults, and slip distribution on the subfaults was estimated. The largest slip, about 3.7 m, was estimated on the subfaults to the west of epicenter, but the average slip is 1.2 m, much smaller than that estimated by the trial and error method (4 m) which compared the maximum tsunami amplitudes (Satake 1989).

The 2004 Indian Ocean tsunami, caused by the Sumatra-Andaman earthquake, was recorded by satellite altimeters, as well as tide gauges. A joint inversion of tsunami waveforms recorded at 12 tide gauge stations and the sea surface heights measured by three satellites indicates that the tsunami source was about 900 km long (Fujii and Satake 2007). The estimated slip distribution (Fig. 8) indicates that the largest slip, about 13–25 m, was located off Sumatra Island and the second largest slip, up to 7 m, near the Nicobar Islands. Inversion of satellite altimeter data alone supports a longer, about 1400 km long, tsunami source (Hirata et al. 2006), but such a model produces much larger tsunami waveforms than observed at Indian tide gauge stations. Inversion of tide gauge data alone does not support tsunami source beneath Andaman Islands (Tanioka et al.



Tsunamis, Inverse Problem of, Fig. 7 Schematic illustration of tsunami waveform inversion method. (a) Inversion for spatial slip distribution (Satake 1989). (b) Inversion for temporal and spatial slip distributions (Satake et al. 2013)

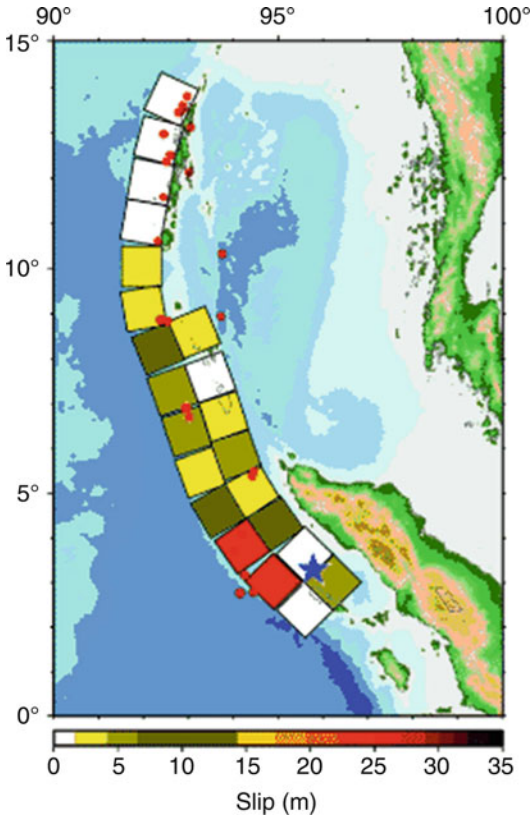
2006). The slip distribution estimated by the joint inversion is similar to those estimated from seismological analyses. The fault slip was the largest near off the northern Sumatra, followed by off Nicobar Islands (Ammon et al. 2005). Fault slip around Andaman Islands was estimated to be small from seismological analysis (Velasco et al. 2006).

For the 2011 Tohoku earthquake tsunami, many cabled bottom pressure gauges and offshore GPS wave gauges recorded the tsunami propagation near the source area. Because these gauges record sea-level change with high sampling interval (~ 1 s), the tsunami waveforms are available at high sampling rates. This enabled to expand the inversion method (Fig. 7b) for temporal change of slip as well as spatial distribution (Satake et al. 2013). More than 50 tsunami waveforms recorded at various types of gauges were used for the analysis. The inversion result (Fig. 9) shows that rupture started near the hypocenter, and the very large (>25 m) slip occurred on the deep plate interface near the hypocenter within ~ 2.5 min, and then

huge (up to 69 m) slip occurred along the trench axis at 3 min after the origin time. This huge shallow slip then propagated to the north with >20 m slip occurred at around 4 min after the rupture initiation. The final slip distribution shows that the slip increases toward the trench axis.

Nonlinear Inversion Methods

In the above inversion method, both tsunami generation and propagation process are assumed to be linear. Because slip amount u , among the nine static fault parameters, is linearly related with the seafloor deformation and tsunami amplitudes, and it has the largest effect on tsunamis, other parameters are fixed in the above method. The tsunami waveforms inversion can be made for water surface displacement without assuming a fault model (Baba et al. 2005; Tsushima et al. 2009). The tsunami propagation, particularly near coasts, might be affected by some nonlinear process such as advection or bottom friction. In addition, for tsunamis generated from relatively small sources or propagating long distance (e.g.,



Tsunamis, Inverse Problem of, Fig. 8 Slip distribution on 22 subfaults of the 2004 Sumatra-Andaman earthquake estimated from a joint inversion of tsunami waveforms on tide gauges and sea surface heights measured by satellite altimeters (Fujii and Satake 2007)

trans-Pacific tsunami), deviation from linear long-wave assumption may become significant. For such a case, the Boussinesq equation including dispersion effects can be used for the numerical simulation, and the computed waveforms are used for the inversion (Saito et al. 2010).

Nonlinear inversion methods have been proposed. Pires and Miranda (2001) proposed an adjoint method, which consists of four steps: source area delimitation by backward ray tracing, the optimization of the initial sea state, nonlinear adjustments of the fault model, and final optimization of fault parameters. The minimum of residual J is obtained iteratively through gradient descent method using the partial derivative or gradient of J with respect to parameters to be inverted.

Inversion of Tsunami Heights

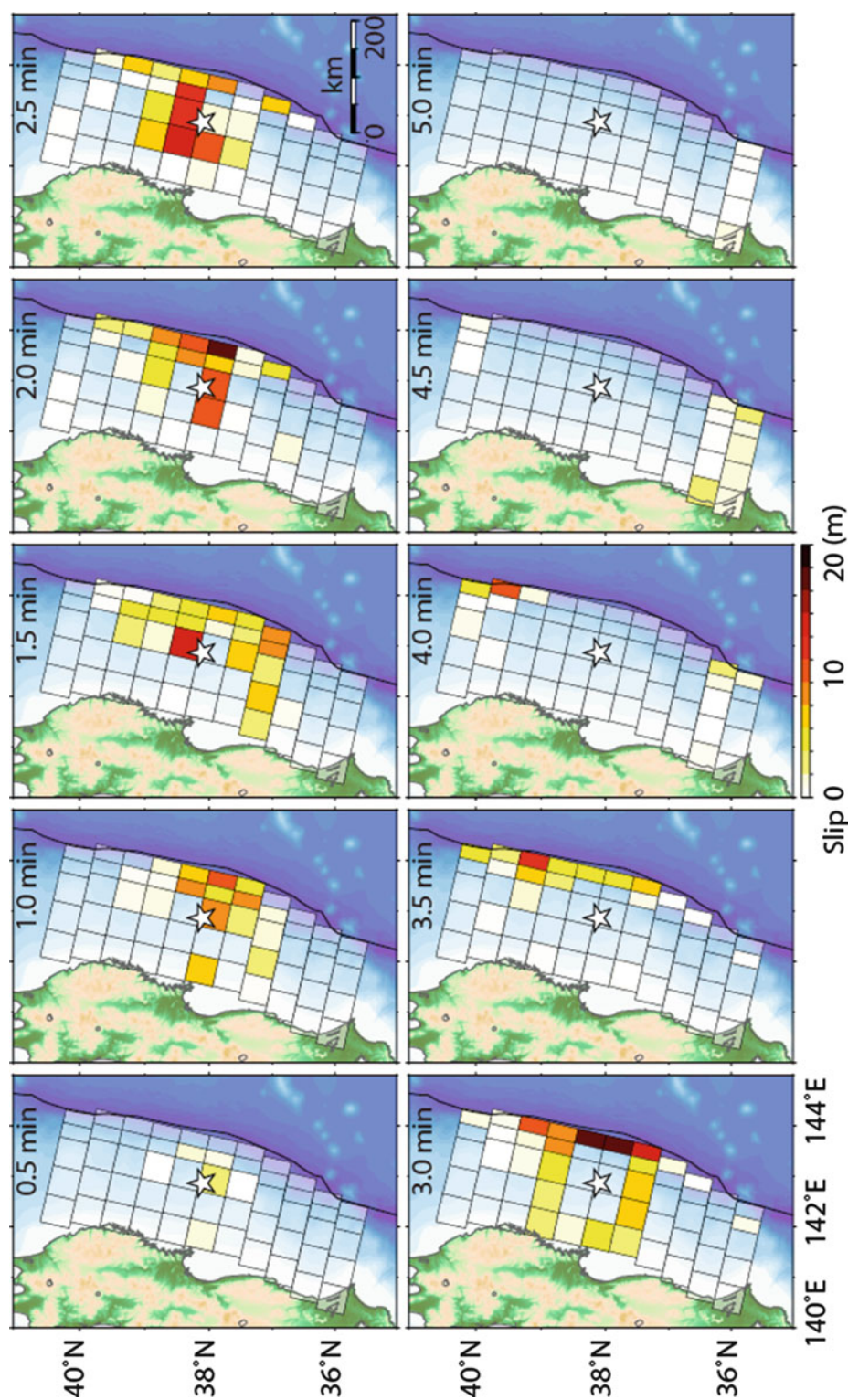
Maximum tsunami heights on coasts, rather than tsunami waveforms, have also been used for tsunami inversion. Distribution of maximum tsunami heights along the coasts is available from field surveys, historical or geological studies, and is valuable to study tsunami source. Piatanesi et al. (1996) used coastal tsunami heights of the 1992 Nicaragua “tsunami earthquake” and estimated the slip distribution on the fault, as well as the mean amplification factor of computed coastal heights and measured run-up heights.

Annaka et al. (1999) proposed a method of joint inversion of tsunami waveforms and run-up heights. As a residual to be minimized, they used a weighted sum of difference in waveforms (similar to Eq. 16) and logarithm of run-up heights. They first tried linear inversion to estimate the initial value and then estimated the perturbation by the nonlinear inversion.

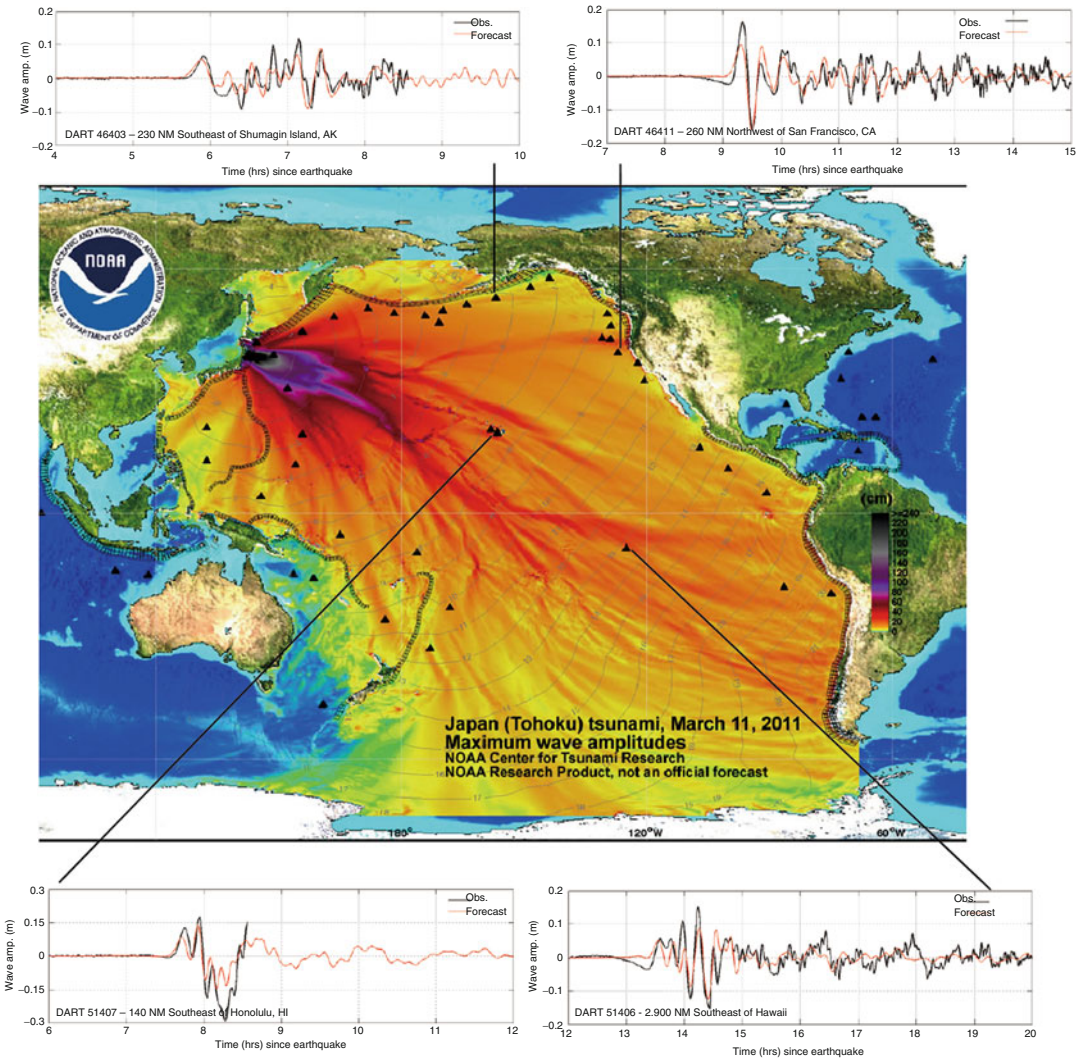
Real-Time Data Assimilation

The tsunami waveform inversion can be done in real time for the purpose of tsunami warning. The real-time data assimilation using the DART records and tsunami forecast has been developed by NOAA (Titov et al. 2005). They first use seismological information to determine the source location and parameters and then, using the database of precomputed simulation results, invert the DART data to estimate the tsunami source size (slip amounts). The tsunami forecast is made for farther locations where the tsunami has not arrived. For the 2011 Tohoku earthquake, they successfully forecasted tsunami waveforms at DART stations based on the data assimilation using the two nearly DART records (Fig. 10).

For near-field tsunamis, Tsushima et al. (2009) developed a similar method, called tsunami Forecasting based on Inversion for initial sea-Surface Height (tFISH), in which the initial sea surface height, rather than slip amount on the prescribed fault, is inverted from tsunami waveforms. They apply this method to tsunami waveforms recorded offshore bottom pressure gauges and demonstrated that it can be used for real-time tsunami warning. Accordingly, Japan Meteorological



Tsunamis, Inverse Problem of, Fig. 9 Snapshot of slip distribution of the 2011 Tohoku earthquake (Satake et al. 2013). The slip distributions were estimated at 0.5 min interval for each of 55 subfaults



Tsunamis, Inverse Problem of, Fig. 10 Real-time data assimilation for the March 11, 2011, Tohoku tsunami performed at NOAA/PMEL/Center for Tsunami Research. *Black triangles* are locations of DART buoys. The tsunami

waveforms recorded on DART (*black curves*) are compared with forecast waveforms (*red curves*) based on the matching at the two nearby stations

Agency adopted this method as one of the follow-up techniques for operational tsunami warning system.

Data assimilation method, such as optimal interpolation or Ensemble Kalman filter technique, has been developed and used in numerical weather forecasting. Maeda et al. (2015) applied the optimal interpolation method for synthesized tsunamis for array of bottom pressure gauge data offshore Japan and demonstrated its ability for

tsunami forecast. The method was further tested using real tsunami data from the 2012 Haida Gwaii earthquake recorded on the ocean bottom seismometer (OBS) network of the Cascadia Initiative in the Cascadia subduction zone (Gusman et al. 2016; Wang et al. 2017). The data assimilation method does not require or assume information on the tsunami source; rather they forecast the tsunami propagation based on the observed data recorded in the network.

Future Directions

Inverse modeling methods of tsunami need to be further developed to better understand the tsunami generation process. Future developments are expected in each field of observation, propagation modeling, and application to seismic and non-seismic tsunamis.

The tsunami observation system has been improved recently, particularly after the 2004 Indian Ocean tsunami. Many instrumental data, both coastal and offshore, become available for the studies of tsunami generation process. Maintenance of the systems, particularly for offshore systems, is sometimes costly, but essential to record infrequent tsunami. Open and real-time availability of such data is also important for tsunami studies as well as for tsunami warning purposes.

For the past tsunamis, more studies are needed to estimate tsunami heights from historical documents, as well as geological data such as distribution of tsunami deposits. Such historical tsunami database has been developed, e.g., at NOAA/NCEI.

For modeling tsunamis recorded on coastal tide gauges or run-up heights, nonlinear computations with very fine bathymetry data are essential. While computational methods have been developed, fine bathymetry data are not always available. Developments of nonlinear inversion methods are also important.

Finally, inversion of tsunami data can be applied to tsunamis generated from submarine processes other than earthquakes, such as volcanic eruptions or landslides. For such nonseismic tsunamis, parametrization is essential to quantify the geological process and to solve inverse problems. In addition, data assimilation technique, which does not assume or require information on tsunami source, can be used for tsunami forecast.

Bibliography

Primary Literature

Abe K (1973) Tsunami and mechanism of great earthquakes. *Phys Earth Planet Inter* 7:143–153

- Abe K (1979) Size of great earthquakes of 1873–1974 inferred from tsunami data. *J Geophys Res* 84:1561–1568
- Abe K (1981) Physical size of tsunamigenic earthquakes of the northwestern Pacific. *Phys Earth Planet Inter* 27:194–205
- Abe K (1989) Quantification of tsunamigenic earthquakes by the Mt scale. *Tectonophysics* 166:27–34
- Aida I (1978) Reliability of a tsunami source model derived from fault parameters. *J Phys Earth* 26:57–73
- Ammon CJ, Ji C, Thio HK, Robinson D, Ni SD, Hjorleifsdottir V, Kanamori H, Lay T, Das S, Helmberger D, Ichinose G, Polet J, Wald D (2005) Rupture process of the 2004 Sumatra-Andaman earthquake. *Science* 308:1133–1139
- Annaka T, Ohta K, Motegi H, Yoshida I, Takao M, Soraoka H (1999) A study on the tsunami inversion method based on shallow water theory. *Proc Coast Eng JSCE* 46:341–345
- Atwater BF, Musumi-Rokkaku S, Satake K, Tsuji Y, Ueda K, Yamaguchi DK (2005) The orphan tsunami of 1700. *USGS Prof Paper* 1707:133
- Baba T, Cummins PR, Hori T (2005) Compound fault rupture during the 2004 off the Kii Peninsula earthquake (M 7.4) inferred from highly resolved coseismic sea-surface deformation. *Earth Planets Space* 57:167–172
- Dawson AG, Shi SZ (2000) Tsunami deposits. *Pure Appl Geophys* 157:875–897
- Fujii Y, Satake K (2007) Tsunami source model of the 2004 Sumatra-Andaman earthquake inferred from tide gauge and satellite data. *Bull Seismol Soc Am* 97: S192–S207
- Fujii Y, Satake K, Sakai S, Shinohara M, Kanazawa T (2011) Tsunami source of the 2011 off the Pacific coast of Tohoku earthquake. *Earth Planets Space* 63:815–820
- Geist E (1998) Local tsunamis and earthquake source parameters. *Adv Geophys* 39:117–209
- Gonzalez FI, Bernard EN, Meinig C, Eble MC, Mofjeld HO, Stalin S (2005) The NTHMP tsunami network. *Nat Hazards* 35:25–39
- Gusman AR, Sheehan AF, Satake K, Heidarzadeh M, Mulia IE, Maeda T (2016) Tsunami data assimilation of Cascadia seafloor pressure gauge records from the 2012 Haida Gwaii earthquake. *Geophys Res Lett* 43:4189–4196
- Hanks T, Kanamori H (1979) A moment magnitude scale. *J Geophys Res* 84:2348–2350
- Hirata K, Satake K, Tanioka Y, Kuragano T, Hasegawa Y, Hayashi Y, Hamada N (2006) The 2004 Indian Ocean tsunami: tsunami source model from satellite altimetry. *Earth Planets Space* 58:195–201
- Imamura F (2009) Tsunami modeling: calculating inundation and hazard maps. In: Bernard EN, Robinson AR (eds) *Tsunamis*. Harvard University Press, Cambridge, pp 321–332
- Intergovernmental Oceanographic Commission (1997) IUGG/IOC TIME project numerical method of tsunami simulation with the leap-frog scheme. UNESCO, Paris.

- http://www.jodc.go.jp/info/ioc_doc/Manual/122367eb.pdf
- Intergovernmental Oceanographic Commission (1998) Post-tsunami survey field guide. UNESCO, Paris
- Kanamori H (1977) The energy release in great earthquakes. *J Geophys Res* 82:2981–2987
- Kato T, Terada Y, Kinoshita M, Kakimoto H, Isshiki H, Matsuishi M, Yokoyama A, Tanno T (2000) Real-time observation of tsunami by RTK-GPS. *Earth Planets Space* 52:841–845
- Kikuchi M, Fukao Y (1985) Iterative deconvolution of complex body waves from great earthquakes – the Tokachi-oki earthquake of 1968. *Phys Earth Planet Inter* 37:235–248
- Lay T, Kanamori H (1981) An asperity model of large earthquake sequences. In: Simpson DW, Richards PG (eds) *Earthquake prediction – an international review*. American Geophysical Union, Washington, DC, pp 579–592
- Lay T, Kanamori H, Ammon CJ, Nettles M, Ward SN, Aster RC, Beck SL, Bilek SL, Brudzinski MR, Butler R, DeShon HR, Ekstrom G, Satake K, Sipkin S (2005) The great Sumatra-Andaman earthquake of 26 December 2004. *Science* 308:1127–1133
- Mader CL (1988) *Numerical modeling of water waves*. University of California Press, Berkeley
- Maeda T, Obara K, Shinohara M, Kanazawa T, Uehira K (2015) Successive estimation of a tsunami wavefield without earthquake source data: a data assimilation approach toward real-time tsunami forecasting. *Geophys Res Lett* 42:7923–7932
- Mansinha L, Smylie DE (1971) The displacement fields of inclined faults. *Bull Seismol Soc Am* 61:1433–1440
- Mei CC (1989) *The applied dynamics of ocean surface waves*. World Scientific, Singapore
- Merrifield MA, Firing YL, Aarup T, Agricole W, Brundrit G, Chang-Seng D, Fare R, Kilonsky B, Knight W, Kong L, Magori C, Manurung P, McCreery C, Mitchell W, Pillay S, Schindele F, Shillington F, Testut L, Wijeratne EMS, Caldwell P, Jardin J, Nakahara S, Porter FY, Turetsky N (2005) Tide gauge observations of the Indian Ocean tsunami, December 26, 2004. *Geophys Res Lett* 32. <https://doi.org/10.1029/2005GL022610>
- Mikada H, Mitsuzawa K, Matsumoto H, Watanabe T, Morita S, Otsuka R, Sugioka H, Baba T, Araki E, Suyehiro K (2006) New discoveries in dynamics of an M8 earthquake-phenomena and their implications from the 2003 Tokachi-oki earthquake using a long term monitoring cabled observatory. *Tectonophysics* 426:95–105
- Miyabe N (1934) An investigation of the Sanriku tsunami based on mareogram data. *Bull Earthquake Res Inst* 1:112–126
- Mori J, Shimazaki K (1985) Inversion of intermediate-period Rayleigh waves for source characteristics of the 1968 Tokachi-oki earthquake. *J Geophys Res* 90:11374–11382
- Murotani S, Satake K, Fujii Y (2013) Scaling relations of seismic moment, rupture area, average slip, and asperity size for $M \sim 9$ subduction-zone earthquakes. *Geophys Res Lett* 40:5070–5074
- Nagarajan B, Suresh I, Sundar D, Sharma R, Lal AK, Neetu S, Shenoi SSC, Shetye SR, Shankar D (2006) The great tsunami of 26 December 2004: a description based on tide-gauge data from the Indian subcontinent and surrounding areas. *Earth Planets Space* 58:211–215
- Nanayama F, Satake K, Furukawa R, Shimokawa K, Atwater BF, Shigeno K, Yamaki S (2003) Unusually large earthquakes inferred from tsunami deposits along the Kuril trench. *Nature* 424:660–663
- Neetu S, Suresh I, Shankar R, Shankar D, Shenoi SSC, Shetye SR, Sundar D, Nagarajan B (2005) Comment on “The great Sumatra-Andaman earthquake of 26 December 2004”. *Science* 310:1431a
- Okada Y (1985) Surface deformation due to shear and tensile faults in a half-space. *Bull Seismol Soc Am* 75:1135–1154
- Piatanesi A, Tinti S, Gavagni I (1996) The slip distribution of the 1992 Nicaragua earthquake from tsunami run-up data. *Geophys Res Lett* 23:37–40
- Pires C, Miranda PMA (2001) Tsunami waveform inversion by adjoint methods. *J Geophys Res* 106:19773–19796
- Rabinovich AB, Thomson RE (2007) The 26 December 2004 Sumatra tsunami: analysis of tide gauge data from the world ocean part 1, Indian Ocean and South Africa. *Pure Appl Geophys* 164:261–308
- Saito T, Satake K, Furumura T (2010) Tsunami waveform inversion including dispersive waves: the 2004 off Kii Peninsula earthquake. *J Geophys Res* 115:B06303. <https://doi.org/10.1029/2009JB006884>
- Satake K (1988) Effects of bathymetry on tsunami propagation – application of ray tracing to tsunamis. *Pure Appl Geophys* 126:27–36
- Satake K (1989) Inversion of tsunami waveforms for the estimation of heterogeneous fault motion of large submarine earthquakes – the 1968 Tokachi-Oki and 1983 Japan Sea earthquakes. *J Geophys Res* 94:5627–5636
- Satake K, Tanioka Y (2003) The July 1998 Papua New Guinea earthquake: mechanism and quantification of unusual tsunami generation. *Pure Appl Geophys* 160:2087–2118
- Satake K, Wang KL, Atwater BF (2003) Fault slip and seismic moment of the 1700 Cascadia earthquake inferred from Japanese tsunami descriptions. *J Geophys Res* 108. <https://doi.org/10.1029/2003JB002521>
- Satake K, Fujii Y, Harada T, Namgaya Y (2013) Time and space distribution of coseismic slip of the 2011 Tohoku earthquake as inferred from tsunami waveform data. *Bull Seismol Soc Am* 103:1473–1492
- Sawai Y, Namegaya Y, Okamura Y, Satake K, Shishikura M (2012) Challenges of anticipating the 2011 Tohoku earthquake and tsunami using coastal geology. *Geophys Res Lett* 39:L21309. <https://doi.org/10.1029/2012GL053692>

- Smith WHF, Scharroo R, Titov VV, Arcas D, Arbic BK (2005) Satellite altimeters measure tsunami, early model estimates confirmed. *Oceanography* 18:10–12
- Stein S, Okal EA (2005) Speed and size of the Sumatra earthquake. *Nature* 434:581–582
- Steketee JA (1958) On Volterra's dislocations in a semi-infinite elastic medium. *Can J Phys* 36:192–205
- Synolakis CE, Okal EA (2005) 1992–2002: perspective on a decade of post-tsunami surveys. In: Satake K (ed) *Tsunamis: case studies and recent developments*. Springer, Dordrecht, pp 1–29
- Tanioka Y, Satake K (1996) Tsunami generation by horizontal displacement of ocean bottom. *Geophys Res Lett* 23:861–864
- Tanioka Y, Yudhicara KT, Kathirolu S, Nishimura Y, Iwasaki S-I, Satake K (2006) Rupture process of the 2004 great Sumatra-Andaman earthquake estimated from tsunami waveforms. *Earth Planets Space* 58:203–209
- Titov VV, Gonzalez FI, Bernard EN, Eble MC, Mofjeld HO, Newman JC, Venturato AJ (2005) Real-time tsunami forecasting: challenges and solutions. *Nat Hazards* 35:41–58
- Tsai VC, Nettles M, Ekstrom G, Dziewonski AM (2005) Multiple CMT source analysis of the 2004 Sumatra earthquake. *Geophys Res Lett* 32. <https://doi.org/10.1029/2005GL023813>
- Tsushima H, Hino R, Fujimoto H, Tanioka Y, Imamura F (2009) Near-field tsunami forecasting from cabled ocean bottom pressure data. *J Geophys Res* 114: B06309
- Velasco AA, Ammon CJ, Lay T (2006) Search for seismic radiation from late slip for the December 26, 2004 Sumatra-Andaman ($M_w = 9.15$) earthquake. *Geophys Res Lett* 33:L18305. <https://doi.org/10.1029/2006GL027286>
- Wang Y, Satake K, Maeda T, Gusman AR (2017) Green's Function-based Tsunami Data Assimilation: a fast data assimilation approach toward tsunami early warning. *Geophys Res Lett* 44:10282–10289
- Yamanaka Y, Kikuchi M (2004) Asperity map along the subduction zone in northeastern Japan inferred from regional seismic data. *J Geophys Res* 109:B07307. <https://doi.org/10.1029/2003JB002683>
- Yamashita T, Sato R (1974) Generation of tsunami by a fault model. *J Phys Earth* 22:415–440
- Yanagisawa K, Imamura F, Sakakiyama T, Annaka T, Takeda T, Shuto N (2007) Tsunami assessment for risk management at nuclear power facilities in Japan. *Pure Appl Geophys* 164:565–576
- Yeh H, Liu P, Synolakis C (1996) Long-wave runup models. World Scientific, Singapore

Books and Reviews

- Lawson CL, Hanson RJ (1974) Solving least squares problems. Prentice-Hall, Englewood Cliffs. (Republished by Society for Industrial and Applied Mathematics, 1995)
- Lay T, Wallace TC (1995) Modern global seismology. Academic, San Diego
- Menke W (1989) Geophysical data analysis: discrete inverse theory (revised edition). Academic, San Diego
- Satake K (2007) Tsunamis. In: Kanamori H (ed) *Treatise on geophysics*, vol 4. Elsevier, Amsterdam

Tsunamis: Bayesian Probabilistic Analysis

Anita Grezio¹, Stefano Lorito², Tom Parsons³ and Jacopo Selva¹

¹Istituto Nazionale di Geofisica e Vulcanologia, Bologna, Italy

²Istituto Nazionale di Geofisica e Vulcanologia, Rome, Italy

³USGS, Menlo Park, CA, USA

Article Outline

Glossary

Definition of the Subject

Introduction

Methodology

Bayesian PTHA

Discussion and Conclusions

Future Directions

Bibliography

Glossary

Aleatory variability In the present context, it is the assumed random variability of the parameters characterizing the future hazardous events or, in other words, the random variability in the model describing the physical system under investigation.

Bayesian statistics An approach to statistics which represents unknown quantities with probability distributions that in one interpretation represent the degree of belief that the unknown quantity takes any particular value. Data are considered fixed and the parameters of distributions representing the state of the world or hypotheses are updated as evidences are collected.

Bias The tendency of a measurement process or statistical estimate to over- or underestimate the value of a population parameter on average.

Conditional probability The probability that an event will occur under the condition or given knowledge that another event occurs.

Conjugacy In Bayesian statistics, the property of parametric families of distributions for prior and likelihood that lead the posterior distribution to be of the same family as the prior distribution.

Completeness It is the extent to which all needed statistics are available. In geophysics, it is usually referred to catalogs of past events, and it refers to the spatiotemporal windows in which virtually no events in a given energetic range are missing.

Epistemic uncertainty The uncertainty deriving from limited knowledge of the physical process, usually treated with alternative models of the same process.

Estimation The process by which we make inferences about a population, based on information obtained from a sample.

Exceedance probability The probability that a given parameter will be larger than a threshold value over a time interval of interest.

Frequentist statistics An approach to statistical reasoning which considers the observed sample to be one realization of repeatable random experiment. The parameters to be estimated are considered to be constants, in contrast with Bayesian statistics where the parameters are treated as random variables.

Inference It is the act of generalizing from the data ("sample") to a larger phenomenon ("population").

Joint probability distribution Describes the simultaneous occurrence of two or more events treated as random variables.

Likelihood function A function of the unknown parameters conditioned on the given fixed observed data, which returns the likelihood that the parameters assume specific values.

Probability density function (PDF) Also known as the density of a continuous random variable, it is a function that describes the relative likelihood that a random variable takes a given value. The probability of the random variable falling within a particular range of values is given by the integral of this variable over the given range. The PDF is *nonnegative* everywhere, and its integral over the entire space is equal to one.

Recurrence interval or average return period It is the average time interval between events of a similar size or intensity.

Run-up It is the maximum topographic height reached by inundation above a reference sea level, usually measured at the horizontal inundation limit during a tsunami event.

Definition of the Subject

Tsunamis are low-frequency high-consequences major natural threats, rare events devastating vast coastal regions near and far from their generation areas. They may be caused by coseismic seafloor motions, subaerial and submarine mass movements, volcanic activities (like explosions, pyroclastic flows, and caldera collapses), meteorological phenomena, and meteorite ocean impacts. The probability of tsunami occurrence and/or impact on a given coast may be treated formally by combining calculations based on empirical observations and on models; this probability can be updated in light of new/independent information. This is the general concept of the Bayesian method applied to tsunami probabilistic hazard analysis, which also provides a direct quantification of forecast uncertainties. This entry presents a critical overview of Bayesian procedures with a primary focus on their appropriate and relevant applicability to tsunami hazard analyses.

Introduction

Bayesian inference is a process of learning from data and from experience (Gelman et al. 2013).

We define prior information as the knowledge before observing new data and posterior information as the understanding after considering new data with the improvement of the prior knowledge due to the evidence in the data. Named after British mathematician the Reverend Thomas Bayes (1701–1761), Bayesian inference is based on statements of conditional probability. Bayes' formula, at the roots of Bayesian methods, was introduced in special cases and was formalized in a posthumous paper presented at the Royal Society of London. Stigler (1983) attributes the principle of the Bayesian inference to Saunderson (1683–1739), a professor of optics who published a large number of mathematics papers. From 1774 to 1812, Laplace studied the general concept of the conditional probability independently, considering the inductive probability by reassessing the prior estimates if new relevant evidence has emerged. Due to the work of Laplace, Bayesian statistics was in common use for practical applications beginning around the late nineteenth to early twentieth centuries.

The awareness of the tsunami threat to coastal communities has been increased worldwide after catastrophic, global-scale tsunamis generated by great megathrust earthquakes in the Indian Ocean (26 December 2004), in Chile (27 February 2010), in Japan (11 March 2011) and by several other seismic events in the last decade (Lay 2015; Lorito et al. 2016). Hence, progressively more intense efforts have been spent on estimating long-term tsunami hazard and risk. Probabilistic tsunami hazard analysis (PTHA) provides a quantitative tool of the tsunami hazard assessments for tsunami risk mitigation plans and for the implementation of tsunami early warning systems (Geist and Parsons 2006; Geist and Lynett 2014; Grezio et al. 2017).

PTHA estimates the probability of exceeding specific tsunami intensities (wave heights, flow-depth, run-up, velocity, etc.; see, e.g., TPSWG 2006) within a certain time period (exposure time) at given locations (key sites). The most common approaches to PTHA are either based on combining source probability with numerical modeling of the ensuing tsunamis (computationally based approach: Annaka et al. 2007;

Burbidge et al. 2008; Davies et al. 2016; Gonzalez et al. 2009; Heidarzadeh and Kijko 2011; Hoechner et al. 2016; Horspool et al. 2014; Knighton and Bastidas 2015; Lane et al. 2013; Lorito et al. 2015; Mueller et al. 2015; Omira et al. 2015; Power et al. 2013; Sakai et al. 2006; Selva et al. 2016; Sørensen et al. 2012; Suppasri et al. 2012; Thio et al. 2010; Thio and Li 2015) or, less frequently, based on the observed tsunami frequency at a given coastal site (empirical approach: Geist et al. 2014; Orfanogiannaki and Papadopoulos 2007; Tinti et al. 2005; Yadav et al. 2013). Empirical methods are often inhibited by a paucity of information. For this reason, computationally based methods are often preferred for tsunamis; however, they can be tremendously computationally intensive, and the statistics and physics of the sources can be difficult to constrain and model (see discussion on the different approaches and their advantages/limitations in Geist and Lynett 2014).

Bayesian statistics allows merging these different approaches, homogeneously integrating the different kinds of information and thus may represent an effective method to assess tsunami hazard. In examples of computationally based versus empirical approaches, the computationally based analyses may constitute the core assessments for building a first attempt of a complete description of the probabilistic tsunami hazard. The computationally based approach covers all the tsunami sources that are considered possible and their (natural) aleatory variability. Direct numerical modeling of the tsunamis generated by each assumed source scenario is then performed. The exceedance probability at a given site is finally assessed by combining the simulation results with the source probability. Alternative (source and propagation) models span the epistemic uncertainty. Furthermore, this prior information may be updated considering the statistics of the observed tsunamis. Bayesian techniques enable the merging of models and observations into a coherent probabilistic framework, and they may be adopted at some stage of the computationally based PTHA, for example, to constrain the earthquake or tsunami magnitude-frequency relationships and associated uncertainties, or the

likelihood of the earthquake focal mechanism at a given location (e.g., Shin et al. 2015; Selva et al. 2016; Yadav et al. 2013), or for the final results of the PTHA (e.g., Grezio et al. 2012; Parsons and Geist 2009).

Bayesian approaches have been applied not only for long-term PTHA but also for short-term and/or time-dependent tsunami hazard forecast, in the context of tsunami early warning (Blaser et al. 2011; Tatsumi et al. 2014).

Here we present the general Bayesian methodology, and two examples of Bayes' theorem applied to the tsunami hazard analysis: (i) developing a tsunami forecast from numerical modeling and an empirical catalog and (ii) implementing weighting factors based on past tsunami data for statistical models estimating run-up exceedance rates and run-up forecast with subjective estimation of the variance. Finally, we discuss the advantages and the limitations of the Bayesian approach in tsunami probabilistic analysis, and we outline some future directions of the Bayesian approach in tsunami studies.

Methodology

In general terms, Bayesian statistics identifies a given number of unknown parameters Θ and tries to infer their values θ by accounting for some measurable data $Y = \{y_1, \dots, y_i, \dots, y_n\}$ and for the knowledge, independent from data, that we may have about their values. In all the steps of the analysis, the uncertainty is expressed through probability density functions (PDFs), and, in Bayesian analyses, all the parameters (including probability values) can be treated as unknowns (Draper 2009; Gelman et al. 2013).

In PTHA, the parameters under investigation are generally the exceedance probabilities of a given tsunami intensity at one site ($\Theta = P(Z > z; x, DT)$), where the tsunami intensity is, for example, the run-up. The data may be observed intensities at given sites due to past tsunamis or the observed frequency of exceedance in a past time interval DT (Grezio et al. 2010). The parameters in question may also be intermediate quantities required for building the PTHA, like earthquake

magnitudes and focal mechanisms ($\Theta = \{M, \text{strike}, \text{dip}, \text{rake}\}$) or landslide volume or shape (Grezio et al. 2012).

The current state of knowledge about the parameters Θ (i.e., the relative ignorance of the parameter values) and the relative uncertainty before considering new observations are expressed through the prior PDF, hereinafter indicated with $P(\Theta)$. This term allows to account for whatever source of information, from theoretical models to expert beliefs and for both quantitative assessments and qualitative information. If no independent information is available, it is possible to set *noninformative* prior distributions (Box and Tiao 1992).

In order to understand what the observables say about the parameters Θ , we have to set a parametric statistical model that links the observables Y to the parameters. Then, the likelihood of the observables, given a specific value of Θ , needs to be evaluated. This information is expressed through the likelihood PDF, hereinafter indicated as $P(Y|\Theta)$, which gives the probability density of the observed data for any choice of model parameters and follows from the assumed parametric statistical model.

The final goal is the quantification of the probability of the different values of the parameters Θ given the prior knowledge and the observations Y . This information is expressed through the posterior PDF, hereinafter indicated as $P(\Theta|Y)$. The posterior distribution is computed from the prior and the likelihood distributions through *Bayes' theorem*, discussed in the following paragraph. In practice, the past data are utilized to compute the likelihood function and then update the prior beliefs. If new evidence is then gathered, such as new data, the described procedure may be applied iteratively, and the old posterior can play the role of the new prior, to be combined with the new likelihood to obtain a new posterior. The results of the Bayesian analysis are conditional on the assumptions on which the statistical models for the prior PDFs and likelihood functions are built. The results allow for multiple interpretations like hypothesis evaluation, inverse probability problem, prediction process, model evaluation,

parameter ranges, and sensitivity analysis (Box and Tiao 1992; Congdon 2006; Gelman et al. 2013).

Bayes' Theorem

Being Θ the model parameters and Y the observed data, *Bayes' theorem* (Gelman et al. 2013) enunciates that the updated posterior probability distribution $P(\Theta | Y)$ is

$$P(\Theta|Y) = \frac{P(\Theta)P(Y|\Theta)}{P(Y)} \quad (1)$$

where:

- the notation $P(\cdot)$ denotes the probability PDF and $P(\cdot|\cdot)$ the conditional PDF, whose parameters are usually referred to as hyper-parameters. Hence, conditional PDFs are central elements in the Bayesian framework.
- $P(\Theta)$ is the prior probability distribution. The Θ parameter spans a range of possible values θ and defines the hypothesis space. The parameters are not estimated as a single point in the parameter space but are instead represented by a distribution and its statistics (e.g., prior mode or prior mean).
- $P(Y|\Theta)$ is the likelihood function. It represents the information about Θ contained in the data Y .
- $P(Y)$ is a normalization constant ensuring posterior probability integrates to 1. For this reason, it is often omitted from the notation, reporting “proportional to” instead of “equal to” in Eq. 1 (Gelman et al. 2013).

The posterior distribution $P(\Theta | Y)$ quantifies the Bayesian inference about the parameters obtained through Bayes' theorem from the prior and likelihood. As for the prior, the parameters are not estimated as a single point in the parameter space but are instead represented by a distribution. The posterior distribution may be seen as an update of the prior (a novel estimate of the same quantity) in light of new data.

Distribution Forms and Mathematical Techniques

The distribution forms of the prior and likelihood should reflect the probabilistic representations of the tsunami parameters and/or statistical descriptions of observed values. In tsunami investigations, probability distributions of a selected parameter Θ commonly make use of:

- *Normal* density ($\Theta \sim \text{Nor}(y | \mu, \sigma)$ where μ is the mean and σ the standard deviation)
- *Poisson* density ($\Theta \sim \text{Pois}(y | \lambda)$ where λ is a constant rate of occurrence in the considered time interval)
- *Gamma* density ($\Theta \sim \text{Gam}(y | \alpha, \beta)$ where the hyper-parameters α and β , respectively, control the probability distribution immediately after each sub-event and describe the longer-term rate)
- *Binomial* density ($\Theta \sim \text{Bin}(y | n, \theta)$ where n are the trials, given that the probability of success in one trial is θ)
- *Beta* density (with $\Theta \sim \text{Beta}(y | \alpha, \beta)$ where the two positive hyper-parameters α and β are the exponents of the random variable)
- *Uniform* density (with $\Theta \sim \text{Unif}(y | \alpha, \beta)$ with all values between α and β equally probable)

Bayesian statistics may utilize the valuable mathematical properties of the conjugate analysis to find prior distributions for the likelihoods (denominated conjugate priors) in order to represent the results in an analytic form which simplifies the computations. When possible, this is obtained by picking a prior distribution of the same “family” of the likelihood function so that the resulting probability distribution is also in the family. In this way, a closed-form expression for the posterior distribution is obtained, and, for example, numerical integration can be avoided. In tsunami problems, like estimating the frequency of events where the rate parameter is unknown but can be constrained somewhat with data, conjugate families are sometimes used. Geist and Parsons (2010) applied a *Poisson-Gamma* conjugate to model the probability of potentially tsunamigenic submarine landslides in the Santa

Barbara Channel (Southern California), Port Valdez (Alaska), and Storegga Slide complex (Norwegian Sea). The landslide probability problem assumes that the landslides are independent and thus occur randomly in time according to a *Poisson* distribution (characterized by a rate parameter λ), and eventually earthquake probability is also considered. Landslide inter-event time uncertainties associated with age dating of individual events and open time intervals were estimated. The seismically imaged landslides typically exhibited only the ages of the youngest and oldest underlying events. However, through the Bayesian approach, even not straightforward information are included. The most likely mean return time ($1/\lambda$) of the submarine landslides was estimated by this *Poisson-Gamma* model using the number of landslide occurrences and the observation period. Grezio et al. (2010) combined the prior *Beta* distribution with the *Binomial* likelihood function in the Messina Strait Area. The posterior distribution is a modified *Beta* distribution constrained by the past run-up observations. In a similar formulation, Knighton and Bastidas (2015) employed the *Poisson-Gamma* conjugate model in the first step to estimate a likelihood function for the *Poisson* rate parameter of tsunamigenic events given in an historical catalog. Then, the likelihood function for the *Poisson* parameter was determined to find the probability of tsunamigenic events causing a hazard exceeding a critical value. The outcome probability was used into a *Beta-Binomial* scheme like in Grezio et al. (2012). Selva et al. (2016) proposed an event tree procedure to quantify source uncertainties in a seismic PTHA. At one level of the event tree, the uncertainty on potential focal mechanisms of earthquakes is modeled. At this node, a *Dirichlet* distribution is used to represent the prior knowledge about the probability of occurrence of the different combinations of discrete intervals of strike, dip, and rake angles. This distribution is then updated by observations of such angles from two earthquake catalogs from the Ionian Sea region (central Mediterranean Sea), naturally distributed following a *Multinomial* distribution. The resulting *posterior* distribution for the probability

of the different interval of angles is again a *Dirichlet* distribution.

If the parameter Θ is a m -dimensional vector $\{\theta_1, \dots, \theta_m\}$, as in the last example of Selva et al. (2016), the probability distributions and the normalization constant $P(Y)$ are m -dimensional problems. If the conjugacy property of prior/likelihood functions is not used to directly obtain the posterior distribution in a closed form, advanced sampling techniques should be adopted. Markov chain Monte Carlo (MCMC) methods are intensive computational techniques used to approximate the high-dimensional integrals associated with the posterior probability distribution in Bayes' theorem. Markov chain samples from the posterior distribution for a time sufficient long to reach the equilibrium within the required approximation (Draper 2009). These techniques extend the range of the single-parameter sampling method to multivariate situations where each parameter or subset of parameters in the overall posterior density may have different density (Congdon 2006). Knighton and Bastidas (2015) evaluate the hazard to the hypothetical coastal facility within a 30-year time period by the Monte Carlo analysis through sampling the likelihood distribution of inter-event timing of the tsunami sources and the beta distribution which pertains to the binomial distribution of the hazard parameter.

Epistemic and Aleatory Uncertainties

All probabilistic analyses typically address the problem of epistemic and aleatory uncertainties. Many authors supported this division, based on either theoretical (Marzocchi and Jordan 2014) or practical (P  -Cornell 1996) reasoning. In its general interpretation, aleatory uncertainty represents the unreducible natural variability of the studied phenomenon, while the epistemic uncertainty arises from the limited knowledge on the system that does not allow to perfectly quantifying the aleatory uncertainty. In this way, it is possible to distinguish the uncertainty that may be reduced by increasing the knowledge of the modeled system (the epistemic uncertainty) from the irreducible unpredictability of the system itself (the aleatory uncertainty). Also, the separation

allows to report the effective variability of the results in a more robust format and to make any probabilistic analysis a testable experiment (Marzocchi and Jordan 2014; Marzocchi et al. 2015).

In probabilistic tsunami hazard analysis, epistemic uncertainty emerges from the substantial lack of understanding of the tsunamigenic processes (e.g., the long-term earthquake rates or the dynamics of "tsunami earthquakes"; see, e.g., Polet and Kanamori 2009) and of the tsunami evolution after generation, or even from approximations in the tsunami numerical modeling made for the sake of practical feasibility (e.g., the common shallow-water approximation), or from the lack of accurate enough digital elevation models.

Both the physics-based and data-driven concepts should address the appropriate hypotheses on the statistical experiment settings. Tsunami events of largest intensity are rare, and the data are often not sufficient to constrain properly the variability of the controlling parameters. As a consequence, a large epistemic uncertainty arises, and many scientifically acceptable alternative models may be formulated.

In the Bayesian paradigm, both types of uncertainty are automatically quantified for the potential reduction of the possible epistemic uncertainty by accounting for all relevant available information. In a Bayesian analysis, the epistemic uncertainties are represented as uncertain parameters, whereas aleatory uncertainties are represented with the choice of probability density functions appearing in the selected parametric statistical model. Since in tsunami applications relatively few data are generally available, parameters are (usually) poorly constrained, and few additional data can feed the likelihood. Thus, the weight of the prior is larger compared to the weight of the likelihood functions.

Advanced approaches tend to extend the exploration of epistemic uncertainty by including alternative statistical models of the aleatory uncertainty, that is, developing the prior probability distributions by implementing different statistical models (Knighton and Bastidas 2015) or implicitly adopting an ensemble modeling approach (Marzocchi et al. 2015; Selva et al. 2016).

Bayesian PTHA

Forecasting tsunamis is typically an underinformed exercise because the mean return time of a given event is often longer than the period we have had to observe it. Thus, we can only rarely develop a complete empirical distribution that satisfactorily captures the aleatory variation. We then rely on numerical models, indirect paleo-evidence, and/or incomplete historical observations to develop probability density function parameters. Many of these models have a stochastic component that attempts to cover the possible range of behaviors. Each of these datasets or model results may capture different aspects of the hazard process. Here we discuss two paradigmatic analyses in which the Bayesian method aggregates a variety of information sources, specifically using likelihood functions shaped from measurement uncertainties and/or stochastic distributions to integrate and weight results. If advanced models, additional and even sparse data, improved instrumental measurements, or new observations become available, an update of the posterior inferences is possible in the Bayesian statistical framework, keeping track of the assumptions on the prior knowledge and the introduced further information.

Example of Tsunami Forecast from Numerical Modeling and an Empirical Catalog

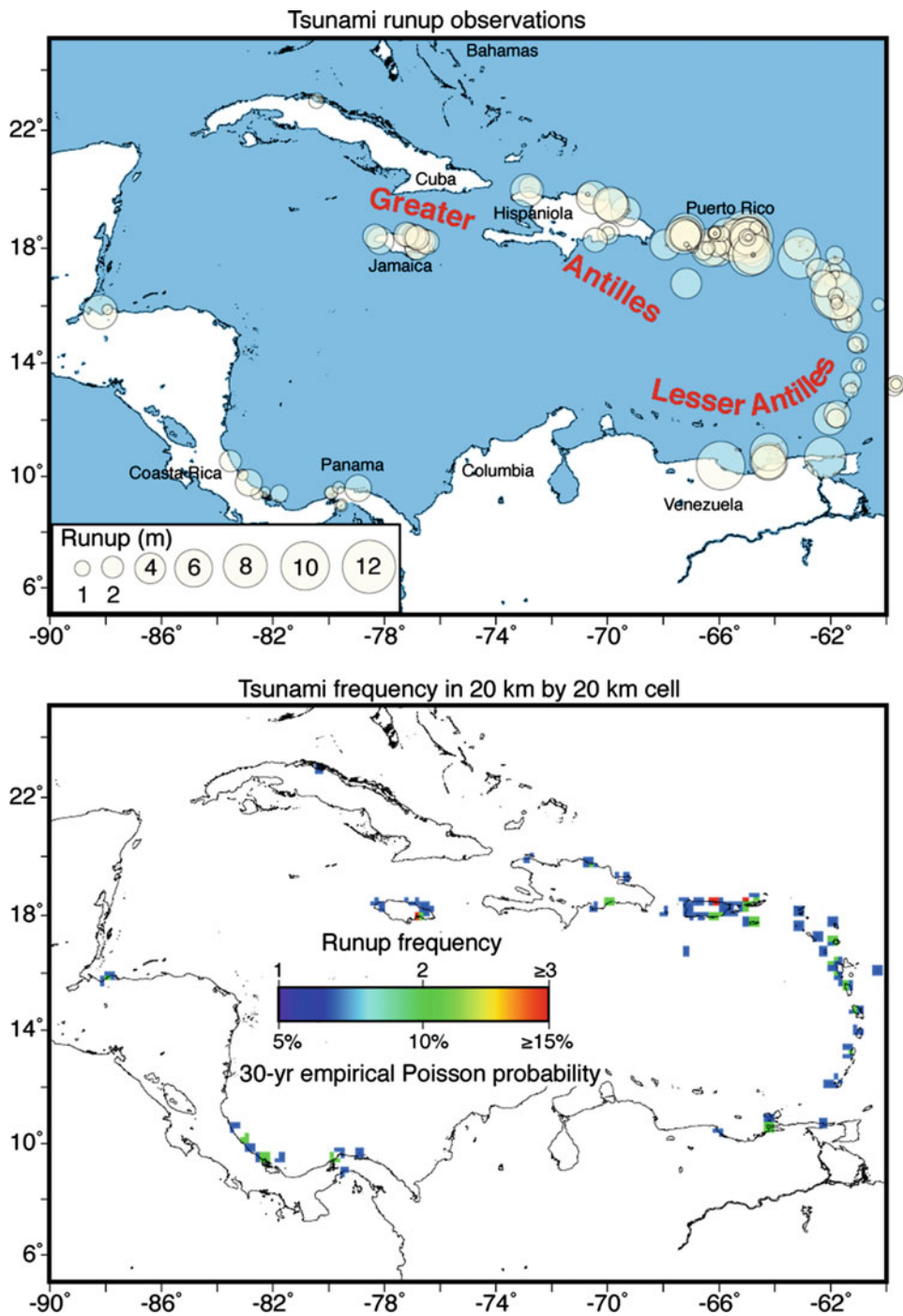
Exceedance rate of run-ups or tsunami waves at a selected site is typically calculated by assessing local and distant tsunamigenic sources, modeling wave height and/or flooding, and then aggregating expected probabilities. It can be difficult to account for every source, especially for the case of submarine landslides. Thus, independent empirical observations, such as from tide gauges, eyewitness accounts, or paleotsunami evidence, are valuable. In this example (after Parsons and Geist 2009), we discuss a spatial tsunami forecast for the Caribbean region (Fig. 1) made from numerical models of wave height (Geist 2002; Geist and Parsons 2006) and an ~500-year-long empirical catalog (O’Loughlin and Lander 2003). Moreover, we discuss how Bayesian methods are used to handle cases where, at a given location,

there may be multiple independent run-up rate distributions derived from different models, only one, or none. We gridded the region uniformly and began with a noninformative prior. In the case we present here, each geographic cell had between zero and two rate distributions that were described by likelihood functions. When there were no estimates for a given cell, then the posterior distribution was zeroed. When one model provided rates, its likelihood function was used to update the priors, and when more than one rate estimate was available, the posterior distribution was developed through combination and renormalization and was then used to update the prior.

The Caribbean region has not produced many large earthquakes in the modern catalog era, which means we have virtually no knowledge of the causative earthquake magnitude-frequency distribution, nor do we have any information about slip distributions. Thus, a wide array of possible earthquake scenarios constrained by the fault geometry and moment rate from plate motions must be modeled. In this case, a 3D finite-element model of the subduction zone and other major faults was constrained by GPS and plate rates/directions to calculate expected fault slip rates, and a set of stochastic earthquake rate models and slip distributions was developed from that. The associated group of 50,500-year simulated catalogs of tsunami run-ups made with a finite-difference approach was then calculated to capture likely variability and modeling uncertainties (see Parsons and Geist 2009 for full details) (Fig. 2). A general aggregation equation for determining the rate (λ) at which tsunamis will exceed a certain run-up (R_0) at a coastal location was used to develop synthetic tsunami catalogs as

$$\lambda(R > R_0) = \sum_{\text{zone}=j} v_j \int_{m_i}^{\infty} P(R > R_0 | m_j) f_j(m) dm. \quad (2)$$

The propagation distance was included in the term $P(R > R_0 | m_j)$ since this term is computed by



Tsunamis: Bayesian Probabilistic Analysis, Fig. 1 *Top* shows individual run-up observations from O’Loughlin and Lander (2003). Circle size represents

run-up in m. *Bottom panel* shows summed number of run-up observations per 20 by 20 km cell and the corresponding empirical *Poisson* probability

numerical propagation models. The moment-frequency distribution for earthquakes in a given zone j is described by the term $f_j(m)$, where a tapered Gutenberg-Richter (G-R) distribution with the complementary cumulative (survivor) distribution $F_j(m)$ of Kagan (2002a) and Kagan and Jackson (2000) was used as

$$F_j(m) = (m_t/m)^\beta \exp\left(\frac{m_t - m}{m_c}\right), \quad m \geq m_t \quad (3)$$

where β is the shape parameter for the distribution, m_t is the threshold moment, and m_c is the corner moment that controls the tail of the distribution. The source rate parameter for each zone (v_j) was defined as the activity rate for earthquakes of $m \geq m_t$ and is related to the seismic moment rate ($\dot{m}_s$) as described by Kagan (2002b) as

$$v(m) = \frac{(1 - \beta)\dot{m}_s}{m^\beta m_c^{1-\beta} \Gamma(2 - \beta) e^{m/m_c}} \quad (4)$$

where Γ is the gamma function. The “tectonic” moment rate ($\dot{m}_t$) is given by $\dot{m}_t = \mu A \dot{u}$, where μ is the shear modulus, A is the area of the seismogenic part of the fault zone, $\dot{u}$ is the long-term slip rate along the fault determined from finite-element modeling, and $\dot{m}_s$ and $\dot{m}_t$ are related by a seismic coupling parameter ($0 \leq c \leq 1$): $\dot{m}_s = c \dot{m}_t$. We implemented Eq. 2 using a *Monte Carlo*-type procedure in which synthetic earthquake catalogs of fixed duration were prepared from random samples of the distribution defined by Eqs. 3 and 4. The primary sources of epistemic uncertainty captured by this process are geographic moment distribution in the form of hypocentral locations, spatial magnitude distribution, and stochastic slip distributions, all of which variably impact the fraction of moment that is tsunamigenic.

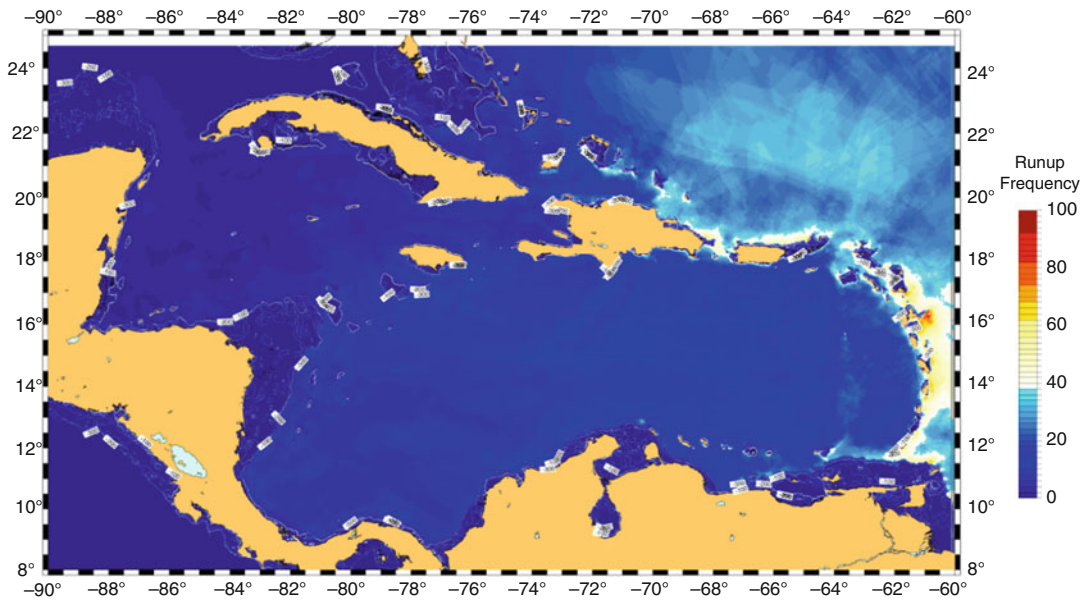
For each earthquake in the synthetic catalogs, vertical and horizontal coseismic seafloor displacements are the initial conditions for tsunami modeling (Tanioka and Satake 1996). Displacements are calculated using Okada’s (1985) analytic functions. A finite rise time of 20 s was

applied uniformly, with no preferred rupture propagation direction. The propagation of the tsunami wavefield is modeled using a finite-difference approximation to the linear long-wave equations (Aida 1969; Satake 2002). A 2 arc minute bathymetric grid (Smith and Sandwell 1997) was used with an 8 s time step that satisfied the *Courant-Friedrichs-Lewy* stability criterion for the Caribbean region. A reflection boundary condition was imposed at the 250 m isobath, whereas a radiation boundary condition was imposed along the open-ocean boundaries of the model (Reid and Bodine 1968). Run-up (R_0) was approximated from the coarse-grid model by finding the nearest model grid point to the coastline and then multiplying the peak offshore tsunami amplitude by a factor of 3 that roughly accounts for shoaling amplification and the run-up process itself (Satake 1995, 2002; Shuto 1991).

We conducted two experiments with a single 4442-year synthetic run-up catalog and another with 50,500-year catalog. We found that the 50,500-year catalogues captured more variability in spatial run-up distribution than did the 4442-year catalogues. This resulted from the multiple catalogs having more varieties of earthquake locations since a few very large events can dominate the distribution of moment, and consequently regional tsunami run-up distribution, due to the Gutenberg-Richter constraint. We thus used the set of 50,500-year catalogues to determine mean rates and uncertainties in the probability calculations.

The empirical catalog, while unusually long, is spatially incomplete because not all coastlines were populated over its duration, meaning that in any given spatial cell, there are less than five observations. Therefore Monte Carlo methods were applied (e.g., Parsons 2008) to extrapolate recurrence parameters (Fig. 3).

It is not uncommon for empirical tsunami rate observations to be higher than numerical models predict in places not fully accounted for by the models. In the Caribbean region, this is seen at Puerto Rico, Jamaica, Costa Rica, and Panama, whereas the numerical rates are higher at the



Tsunamis: Bayesian Probabilistic Analysis, Fig. 2 Example calculation of expected run-up (≥ 0.5 m) frequency over a 4442-year period calculated from the expected seismic moment rate

Antilles (Fig. 4). In these instances, it is likely that the empirical model has captured localized tsunami events that were caused by landslides and/or accommodating faults associated with the plate boundary that were not specifically included in the numerical model sources. Many of the secondary earthquake sources not included have very slow and uncertain slip rates, making implementation into a numerical model difficult. Here, we discuss how these disparate and incomplete distributions can be combined and weighted using likelihood functions.

The primary sources of epistemic uncertainty include (1) tsunami sources not explicitly known or included in the model, (2) seismic coupling coefficient of the Caribbean plate boundary zones, and (3) the degree of completeness in the empirical tsunami catalog. To encompass these uncertainties into probability estimates, a Bayesian framework is created to build tsunami run-up rate estimates within 20 by 20 km cells that contain coastlines throughout the Caribbean region. The key advantage of this approach is that the empirical and model results end up being combined and weighted by their attendant uncertainties.

Having independent empirical and model-derived rate estimates in each spatial cell enables some of the run-up-rate uncertainty to be addressed. *Monte Carlo* fitting of empirical intervals as shown in Fig. 3 along with results from 50 numerical model runs (e.g., Fig. 2) provides arrays of possible run-up rate values at each cell. Unknown/unaccounted-for tsunami sources can be partly accounted for because some of the empirical rates result from sources not accounted for in the numerical model (the most affected areas can be seen by comparing the panels of Fig. 4); the forecast may suffer from incomplete knowledge if events not covered by numerical models have also not occurred in the empirical catalog over the past 500 years. Seismic coupling is a difficult parameter to estimate with certainty; a broad range is captured because the historic earthquake catalog implies a low coupling value of 0.32 (found by comparing seismic moment release to expected slip on Caribbean faults), whereas the numerical models have coupling coefficients of 1.0. Completeness is addressed because low-rate plate-boundary events potentially not seen in the empirical catalog are accounted for with the 50 numerical model runs.

Tsunamis: Bayesian Probabilistic Analysis, Table 1 A 30-year probability of tsunami run-up in excess of 0.5 m in cells that contain population concentrations in 20 by 20 km cells for representative Caribbean countries and territories. Population given as a relative

measure of risk throughout the region. Values were calculated as uniform over cell areas and are not intended to convey any detail at selected cities but are presented for comparison purposes. Dashes indicate negligible calculated probability

Country	Nearest coastal city	Latitude	Longitude	Population	30-yr probability
					$r \geq 0.5$ m (%)
Antigua and Barbuda	St. John's	17.1167°	−61.8500°	24,226	5.74
Belize	Belize City	17.4847°	−88.1833°	70,800	–
Cayman Islands	George Town	19.3034°	−81.3863°	20,626	10.79
Columbia	Cartagena	10.4000°	−75.5000°	895,400	0.08
Costa Rica	Puerto Limon	10.000°	−83.0300°	78,909	8.32
Cuba	Santiago de Cuba	20.0198°	−75.8139°	494,337	2.31
Dominica	Roseau	15.3000°	−61.3833°	14,847	11.94
Dominican Republic	Santo Domingo	18.5000°	−69.9833°	913,540	17.56
France, Guadeloupe	Basse-Terre	16.2480°	−61.5430°	44,864	11.79
France, Martinique	Fort-de-France	14.5833°	−61.0667°	94,049	5.33
Grenada	St. George's	12.0500°	−61.7500°	7500	2.48
Guatemala	Puerto Barrios	15.7308°	−88.5833°	40,900	–
Haiti	Port-au-Prince	18.5333°	−72.3333°	1,277,000	0.01
Honduras	La Ceiba	15.7667°	−86.8333°	250,000	–
Jamaica	Kingston	17.9833°	−76.8000°	660,000	21.95
Netherlands Antilles	Willemstad	12.1167°	−68.9333°	125,000	7.04
Nicaragua	Bluefields	12.0000°	−83.7500°	45,547	–
Panama	Colon	9.3333°	−79.9000°	204,000	17.56
St. Kitts and Nevis	Basseterre	17.3000°	−62.7333°	15,500	6.95
St. Lucia	Castries	14.0167°	−60.9833°	10,634	5.52
St. Vincent and the Grenadines	Kingstown	13.1667°	−61.2333°	25,307	11.32
Trinidad and Tobago	Port of Spain	10.6667°	−61.5167°	49,031	–
Turks and Caicos	Cockburn Town	21.4590°	−71.1390°	5567	3.57
UK, Virgin Islands	Road Town	18.4333°	−64.5000°	9400	13.85
USA, Puerto Rico	San Juan	18.4500°	−66.0667°	434,374	22.24
USA, Virgin Islands	Charlotte Amalie	18.3500°	−64.9500°	18,914	17.56
Venezuela	Cunana	10.4564°	−64.1675°	305,000	6.27

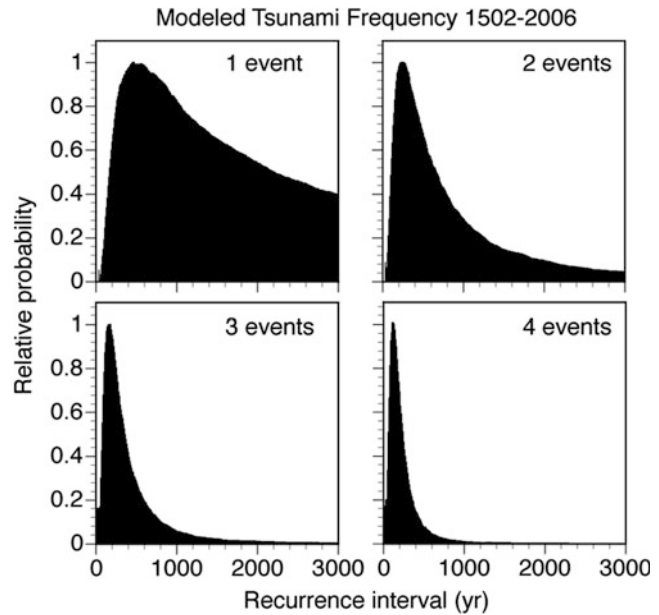
Model-derived run-up rates are combined with empirical rates in the following way: in cells where there are no empirical values, the numerical-model-derived rates are given full weight. Conversely, empirical rates are given full weight where numerical model rates are zero. Lastly, where there are empirical and model rate estimates within the same cells, likelihood functions are used to weight the two models. Distributions shown in Fig. 3 give the relative probability of different rates for a Poisson model that could have caused the empirical observations. Similarly, results from the 50 numerical model runs produce

relative probability (Fig. 5) of different rates in each model cell.

To rank different rate models for each cell where more than one estimate exists (e.g., modeled and observed), a likelihood calculation is made to weight the models. In the simplest, binomial case, likelihood is defined as proportional to the probability of obtaining results A given a fixed hypothesis H resulting from a set of fixed data (equivalent to the sampling distribution as defined in section “[Methodology](#)”). If A_1 and A_2 are two possible, mutually exclusive results, then

Tsunamis: Bayesian Probabilistic Analysis,

Fig. 3 Normalized histograms of the Monte Carlo sequences that matched the indicated event frequencies over 500-year intervals. Ranges of exponential rate parameters are shown (expressed as the inverse, which is recurrence interval) that can match observed frequencies of Caribbean tsunami run-ups (≥ 0.5 m), which range from 1 to 4 events in ~ 500 years



$$P(A_1 \text{ or } A_2 | H) = P(A_1 | H) + P(A_2 | H), \quad (5)$$

$$L(\lambda | e_1, e_2) = k[p_1(e_1 | \lambda)][p_2(e_2 | \lambda)], \quad (7)$$

and likelihood of a specific outcome $A|H$ is defined as its probability; thus,

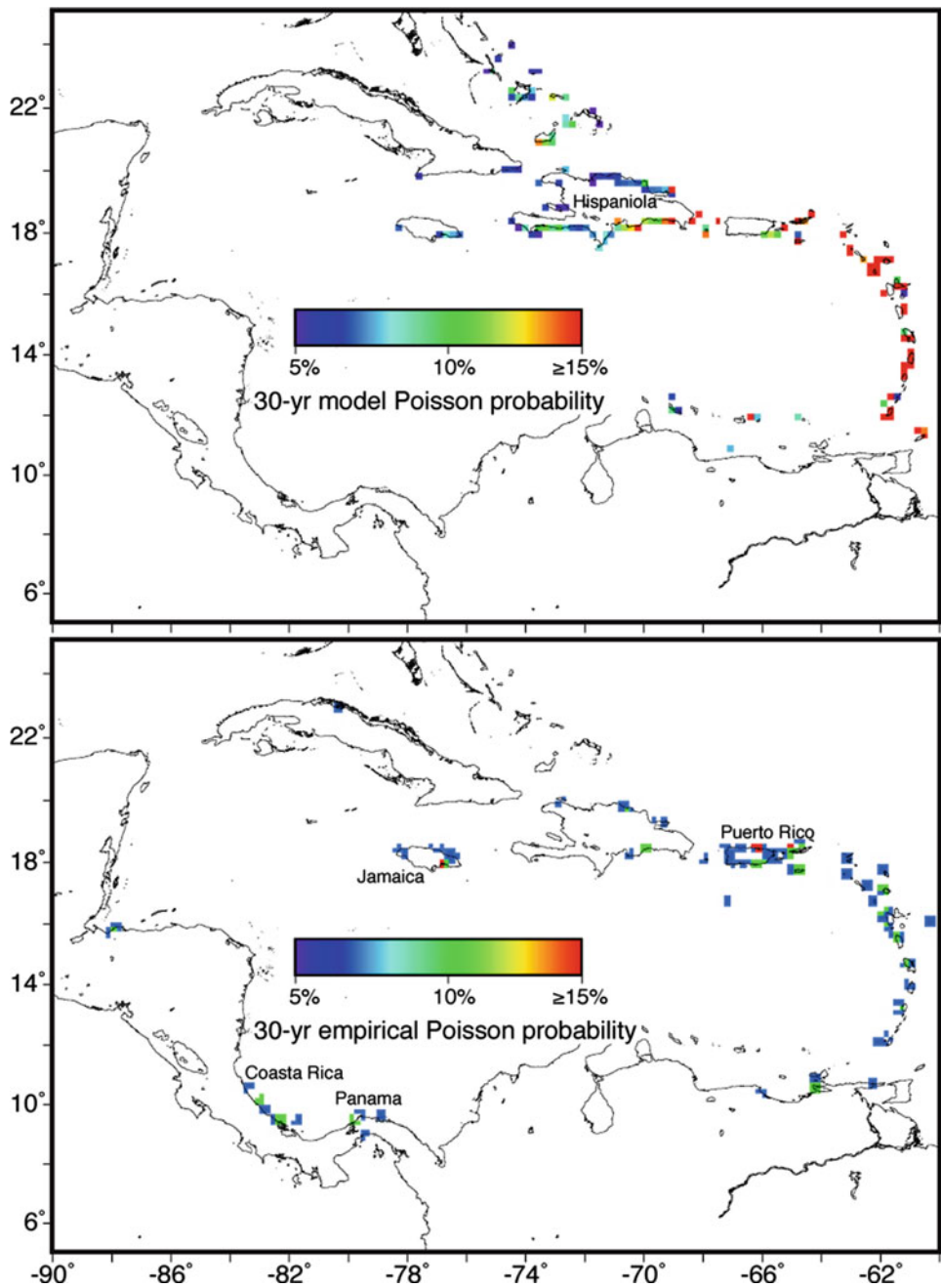
$$L(H|A) = kP(A|H), \quad (6)$$

where k is an arbitrary constant. In the current example, $A_1|H$ is a spatial distribution of numerical model run-up rates, each of which might be correct, whereas $A_2|H$ is a spatial distribution of Monte Carlo-modeled rates based on direct observations.

The results from likelihood functions are used to obtain the final weights using Bayes' rule (Eq. 4), where the posterior distribution is proportional to the likelihood function multiplied by the prior. For this example, we begin with a uniform (noninformative) prior which assumes there is equal probability of all rates in each coastal 20 km by 20 km cell. Next, we update the prior with the empirical results and the numerical model results. Since the prior is updated twice, the same result is achieved by simply multiplying the two likelihood functions. Thus, the likelihood of a given rate λ where there were empirical estimates (e_1) and numerical-modeled estimates (e_2) is

where $p(e_1 | \lambda)$ is the probability of rate λ based on the Monte Carlo fits shown in Fig. 3 and $p(e_2 | \lambda)$ is the probability of rate λ from the 50 numerical model runs. The constant k is used for normalizing the weights so that they add to 1. This can be expanded indefinitely if there are more information sources.

Likelihood functions are used to weight rate models over a range from 0 to 10 events in the 500-year observation period. Rates between 0 and 10 events in 500 years are considered for all cells, assuming no further prior information. Final rates are found by weighted means of the posterior rates. To summarize the process, where model and empirical values are both absent for a given rate, the posterior distribution was zeroed. When one model provides rates, its likelihood function was used to update the priors, and when both empirical and numerical rate estimates are available, likelihood is developed through combination and renormalization using Eq. 4, which is then used to update the priors. Combining empirical and modeled rates makes up for some of the deficiencies in each approach; the



Tsunamis: Bayesian Probabilistic Analysis, Fig. 4 Comparison between (*top*) model-derived 30-year *Poisson* probability (calculated from modeled or observed

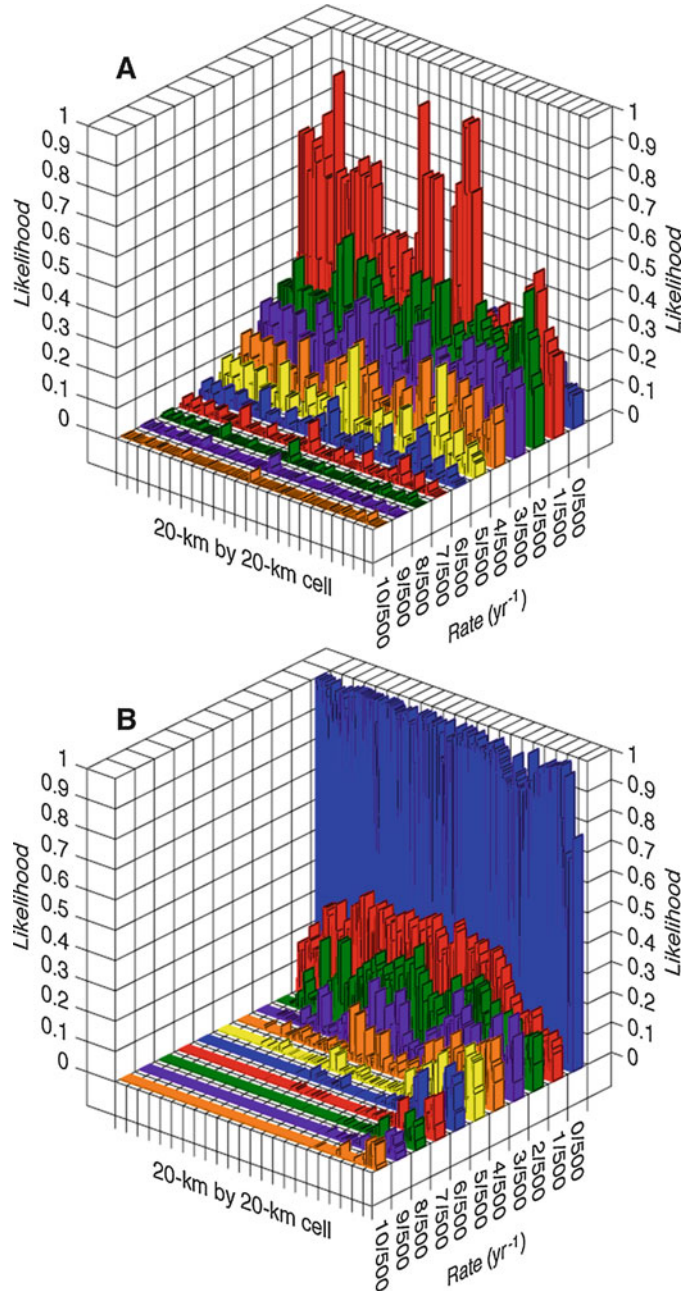
rates in 20×20 km cells) of tsunami run-ups (≥ 0.5 m) and (*bottom*) empirically derived values

empirical catalog is likely not a complete record of all possible interplate tsunami sources, whereas the numerical model did not

account for accommodating intraplate faults and/or landslide sources that appear likely causes of tsunamis in the empirical record.

Tsunamis: Bayesian Probabilistic Analysis,

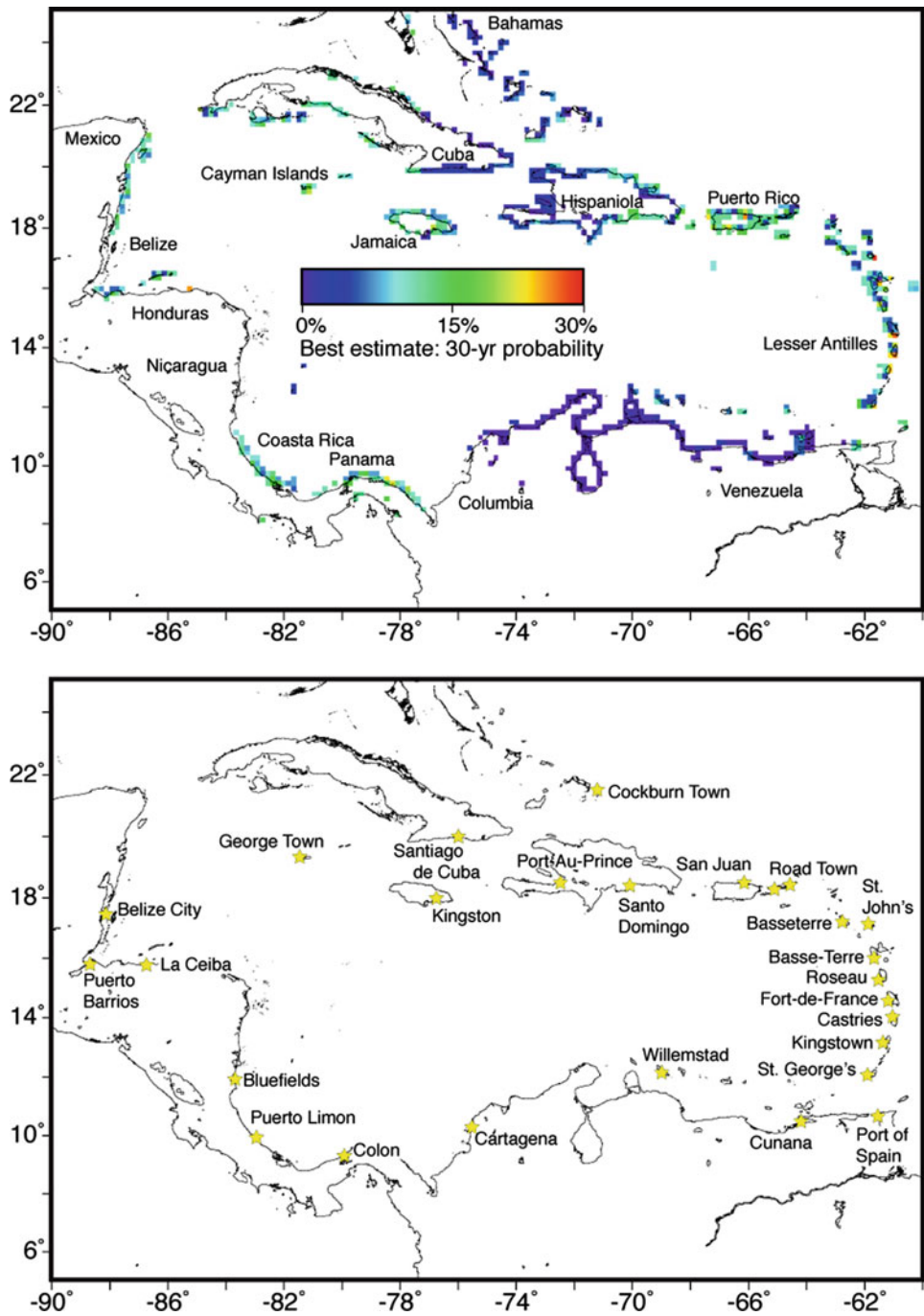
Fig. 5 (a) Normalized histogram (likelihood) of tsunami run-up ($R_0 \geq 0.5$ m) rates in 214 20 km by 20 km cells defined using likelihood functions from empirical rates (Fig. 3) and from 50 numerical modeling simulations. (b) Normalized histogram of run-up rates from numerical modeling in 685 cells where there are no empirical observations. Mean values from these distributions are used in the best-estimate probability calculations mapped in Fig. 6



Example of Past Data Weighting Factors for Statistical Models and Subjective Estimation of the Variance

The Bayesian inference can be applied to the probability Θ that a destructive event overcomes a defined specific threshold z_i of a selected physical parameter Z within a given time period set equal to 1 year. In this example for the Messina

Strait Area, Southern Italy (Fig. 7), the Bayesian PTHA is calculated as the annual probability that the run-up Z overcomes a selected threshold run-up z_i at important sites at least once. Run-up is chosen as intensity as there are historical run-up data reported by the Italian Tsunami Catalogue (Tinti et al. 2004) and other studies (full details in Grezio et al. 2012). In the past centuries, this



Tsunamis: Bayesian Probabilistic Analysis, Fig. 6 A 30-year tsunami run-up ($r \geq 0.5$ m) probability in 20 by 20 km cells at coastal sites in the Caribbean region made

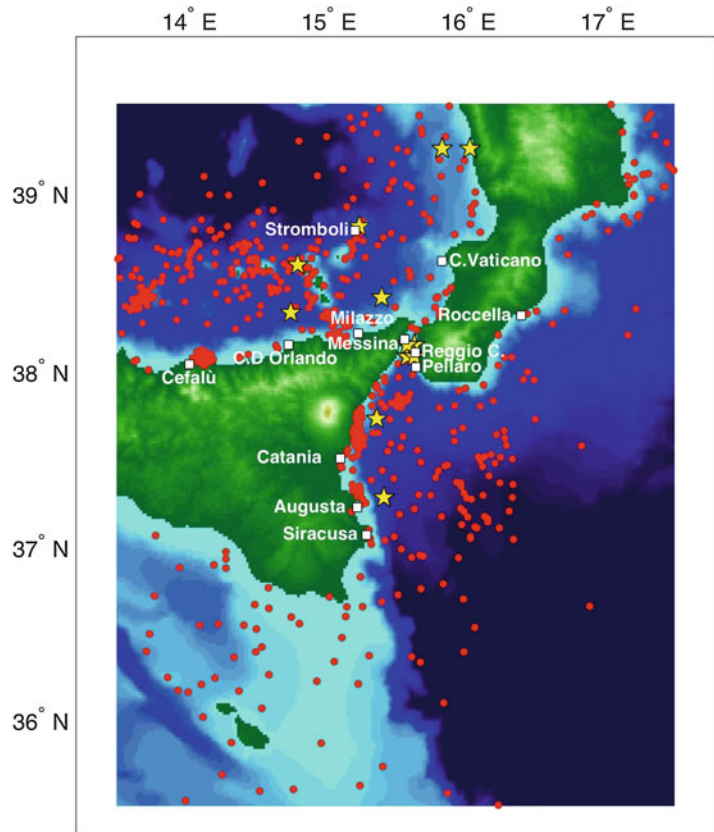
from combined rate estimates from empirical and numerical models. *Lower panel* shows locations of major cities listed in Table 1

area was struck by important tsunamis generated by both regional seismic sources and by non-seismic sources (landslides and total or partial

collapse of volcanic edifices due to volcanic eruptions). Information from the regional seismotectonic studies, marine geology

Tsunamis: Bayesian Probabilistic Analysis,

Fig. 7 Messina Strait Area: dots are the *SSS* epicenters and stars the *SMFs*; the major cities are indicated by squares on the coast where past tsunami events occurred



background, and recent instrumental records are integrated in order to identify the potential tsunamigenic events generated by both submarine seismic sources (*SSSs*) and submarine mass failures (*SMFs*) (Grezio et al. 2010, 2012), and historical indications are used for selecting the key sites. In this example, the prior PDF is derived by (simplified) simulation of the tsunamis generated by *SSSs* and *SMFs*. In the area, these multiple sources are considered as the predominant tsunamigenic sources and are examined to reduce biases and underestimations in the hazard that possibly arise by assuming only one single type of source as primary (Grezio et al. 2015). The likelihood function is used for modeling the past tsunami run-up data. The posterior probability distribution summarizes the updated estimate of the parameter Z .

The *SSSs* are localized on active faults, and the relative epicenters are extracted from the

instrumental Catalogue of the Italian Seismicity with a completeness magnitude of 2.5 (Castello et al. 2007) at depths smaller than 15 km within the shallow part of the crust. Instrumental magnitudes are recorded since 1981, and no tsunami occurred in this short time. In order to consider a large set of potentially tsunamigenic *SSSs*, magnitudes in the range 5.5–7.5 M_w were introduced consistently with the regional seismotectonic studies and weighted using the *Gutenberg-Richter* distributions (Gutenberg and Richter 1944). Finally the magnitudes are associated with the catalog epicenters, and the seafloor deformations are calculated via the analytical formulas by Okada (1992) in order to compute the initial tsunami sea surface waves. The relative fault parameters (width, length, and slip) and focal mechanisms (strike, dip, and rake) are provided, respectively, by the empirical relationships in Wells and Coppersmith (1994) and the

Earthquake Mechanisms of the Mediterranean Area database (Vannucci and Gasperini 2004). The *SSS* spatial distribution is considered uniform.

The *SMFs* are spatially identified using marine geology background knowledge. Their propensity to fail is evaluated on the basis of the mean slope and mean depth, and it is associated to bathymetry cells. In each cell, potentially tsunamigenic *SMFs* are simulated with volumes spanning from 5×10^5 to $5 \times 10^{10} \text{ m}^3$ as indicated by the historical *SMF* sizes identified in the Tyrrhenian and Ionian basins. Additionally, spatial conditional probabilities are introduced considering that the past *SMF* scars represent instability areas. The other geometric parameters and the initial tsunami waves are estimated, respectively, by the rigid body approximation and the empirical formulas in Grilli and Watts (2005) and Watts et al. (2005). In analogy with the subaerial mass failures, the *SMF* frequency-size relationship is assumed to be a power law.

The run-ups Z caused by the *SSS* and *SMF* tsunamigenic sources were calculated through empirical formulas (Synolakis 1987). Uncertainties related to the Z parameter would be reduced through the modeling of source directivity (tsunami energy is not spread isotropically around the source), wave propagation effects (refraction, diffraction, etc.), and other, also non-linear processes during shoaling and coastal inundation (wave breaking, bores, friction, etc.).

The prior probability model of the parameter Z encompasses the theoretical assumptions (e.g., the tsunami modeling), the background knowledge (e.g., the *SMF* spatial distribution and their propensity to failure and the digital elevation model), and the instrumental data on the sources (e.g., the historical seismicity); it does not include the historical tsunami events (which enter in the likelihood distribution). Physical and statistical considerations are made assuming the *Beta* distribution is an adequate subjective choice of the functional form of the prior. Also, the *Beta* distributions are convenient conjugate families for the *Binomial* distribution used for the likelihood, which simplifies the calculations. Assuming that the probability Θ does not vary

in the time interval, the prior distribution for Θ is approximated by a *Beta* distribution with positive α and β in each key site in the Messina Strait Area

$$P(\Theta)_{\text{prior}} \sim \text{Beta}(\alpha, \beta). \quad (8)$$

To choose values of α and β for the prior, we find it easier to work with the expected value E and the variance V for the *Beta* distribution, which are, respectively,

$$\begin{aligned} E(\Theta) &= \frac{\alpha}{\alpha + \beta} \quad \text{and} \\ V(\Theta) &= \frac{E(\Theta)(1 - E(\Theta))}{\alpha + \beta + 1}. \end{aligned} \quad (9)$$

The prior mean E is set equal to the weighted percentage of run-ups $Z > z_t$ generated by the simulated potential tsunamigenic sources

$$E(\Theta) = \sum_i p_i H(Z > z_t) \quad (10)$$

where H is the Heaviside function equal to 1 if the simulated run-up Z is higher than the threshold ($z_t = 0.5 \text{ m}$) and 0 otherwise and p_i is the probability of occurrence of each i -th tsunamigenic source in a time window of 1 year.

The variance V is defined as the *confidence degree* of the prior information through the equivalent number of data Λ ($=\alpha + \beta - 1$) (Marzocchi et al. 2004, 2008; Grezio et al. 2010)

$$V(\Theta) = \frac{E(\Theta)(1 - E(\Theta))}{\Lambda + 2} \quad (11)$$

By setting the parameter Λ to specific values, we assign both the subjective reliability to the prior model and the relative confidence interval. The parameter Λ weights the prior model and represents an estimate of the epistemic uncertainties due to the limited knowledge of the process. In general, a large Λ value corresponds to a large reliability of the prior model, so that the prior distribution needs a great number of past data or observations in the likelihood to be modified significantly. On the contrary, Λ must be small if the

prior model is only a first-order approximation of the process, so that even a limited number of observations in the likelihood can heavily modify the prior distribution. The minimum possible value of A is 1, representing the maximum possible epistemic uncertainty or maximum of level of ignorance. As A increases, the *Beta* function becomes more and more spiked around the given mean. The end-member is a *Dirac's* function judging the epistemic uncertainty negligible when a large amount of data is available (Marzocchi and Lombardi 2008). Here, A is assumed equal to 10 on the basis of practical and expert judgement; it means that more than 10 real data can change drastically the prior probability distribution (Grezio et al. 2012). After computing the expected value E and the variance V , the α and β hyper-parameters of the *Beta* distribution are finally constrained in Eq. 6, and the prior PDF is determined.

The prior *Beta* PDFs for *SSSs* and *SMFs* are shown separately in Fig. 8 for each key site along the Eastern Sicily and the Southern Calabria coasts (Messina Strait Area). The prior probability that a tsunami event overcomes the 0.5 m threshold at each key site is in the interval $[2.2 \times 10^{-4} - 6.4 \times 10^{-4}] \times \text{year}^{-1}$ for the *SSSs* and $[0.7 \times 10^{-7} - 1.5 \times 10^{-7}] \times \text{year}^{-1}$ for the *SMFs*. The relative prior variance is in the range $[1.8 \times 10^{-5} - 5.3 \times 10^{-5}] \times \text{year}^{-1}$ for the *SSSs* and $[0.5 \times 10^{-8} - 1.3 \times 10^{-8}] \times \text{year}^{-1}$ for the *SMFs*.

The likelihood is based on the historical and/or instrumental past data of the run-ups occurred in the Sicily and Calabria regions in the last 500 years (Maramai et al. 2005a,b; Favalli et al. 2009; Tinti et al. 2007; Grezio et al. 2012). The events are assumed independent and investigated in the historical time record of the last 500 years. The set of observations consist of the total number of 1 year time windows in which $Z > z_t$ in the historical catalog. The n years when the tsunami occurred are counted as successes and the years $(n-y)$ without tsunami data as failures. This is formalized with the likelihood function by using a *Binomial* model

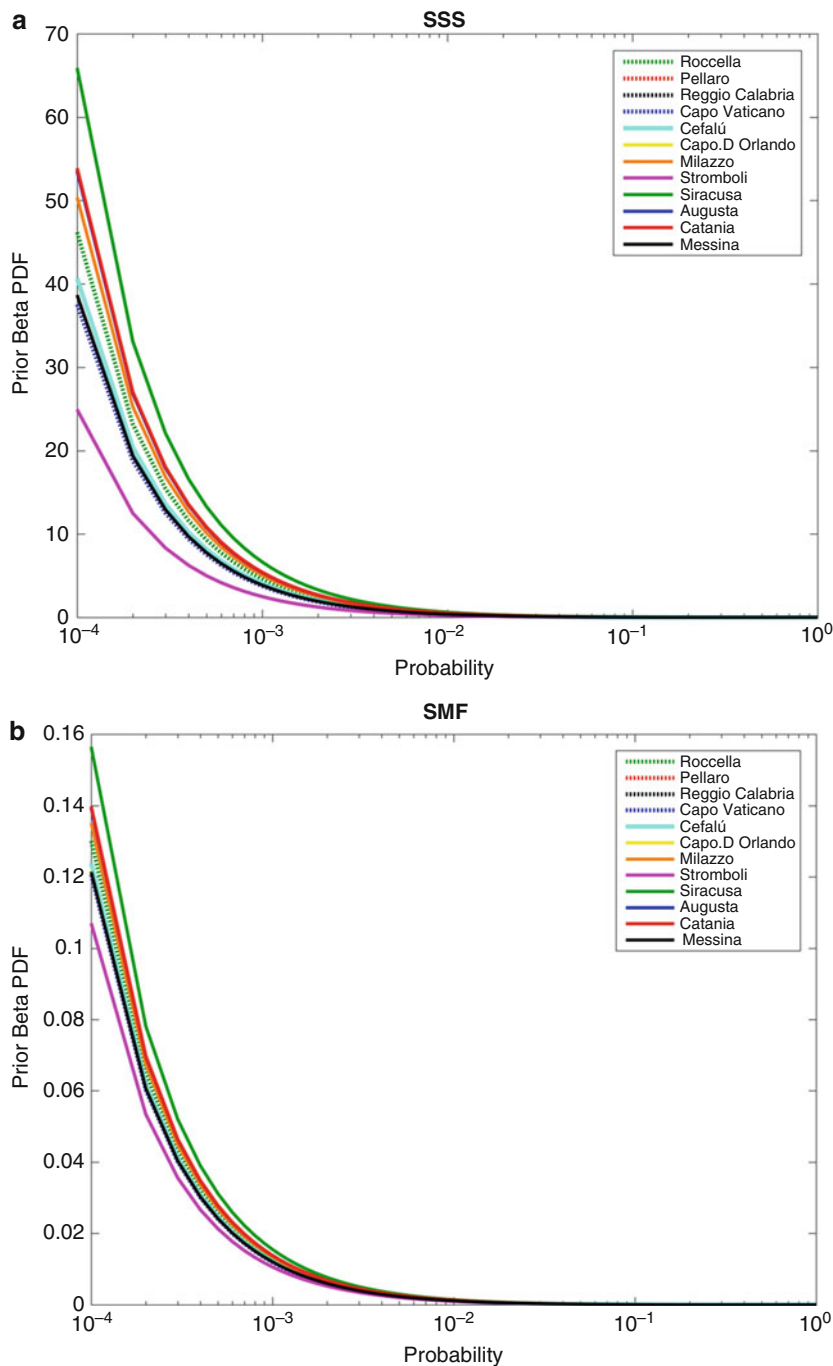
$$P(Y|\Theta = \theta)_{\text{likel}} \sim \text{Bin}(n, \theta) \quad (12)$$

where θ is the random variable defined in the interval $[0, 1]$. From the catalog, we only use entries that have (i) the tsunami reliability equal to 4, meaning that a definite tsunami occurred, and (ii) the tsunami intensity equal to 2 or 3 in the *Ambraseys-Sieberg* scale, recognizing that an event of intensity equal 3 generally produces run-ups of approximately 1 m (Tinti et al. 2005). The impact of the tsunami waves was assumed large in the case of intensity 3, reaching all key sites in the Messina Strait Area, and lower in the case of intensity 2, relevant only for the closest key sites indicated by the catalog. In this case, even if the catalog does not provide explicitly the run-up measures, the values higher than 0.5 m were assigned to the local key sites.

The prior distribution (*Beta* distribution based on physical models, background knowledge, and marine geological information) is modified by the likelihood (*Binomial* distribution based on historical records) producing the posterior distribution (*Beta* distribution computed by point-wise multiplication). Then, the posterior distribution is

$$P(\Theta^i | (Y = y))_{\text{post}} \sim \text{Beta}(a + y^i, \beta + n - y^i) \quad (13)$$

where n is the number of years and y^i the past observed events, with $i = \text{SSS}, \text{SMF}$. The posterior *Beta* PDFs are shown separately in Fig. 9 for each key site. When the historical run-ups are considered in the cities of the Messina Strait Area (Messina, Reggio Calabria, Pellaro, Catania, Augusta, Siracusa, Milazzo, Capo d'Orlando, Cefalù, Capo Vaticano, and Roccella) the posterior means largely increase with values between $2.0 \times 10^{-3} \times \text{year}^{-1}$ and $7.9 \times 10^{-3} \times \text{year}^{-1}$ in the case of the *SSSs*, and at the same time, the variances are reduced at the order $10^{-6} \times \text{year}^{-1}$. Similarly, the posterior means for the *SMFs* increase in Messina, Reggio Calabria, Pellaro, Catania, Augusta, Siracusa, and Stromboli, which experienced mass failures producing tsunamis, with values in the interval $[2.0 \times 10^{-3} - 11.9 \times 10^{-3}] \times \text{year}^{-1}$ and posterior variances of the order $10^{-10} \times \text{year}^{-1}$.

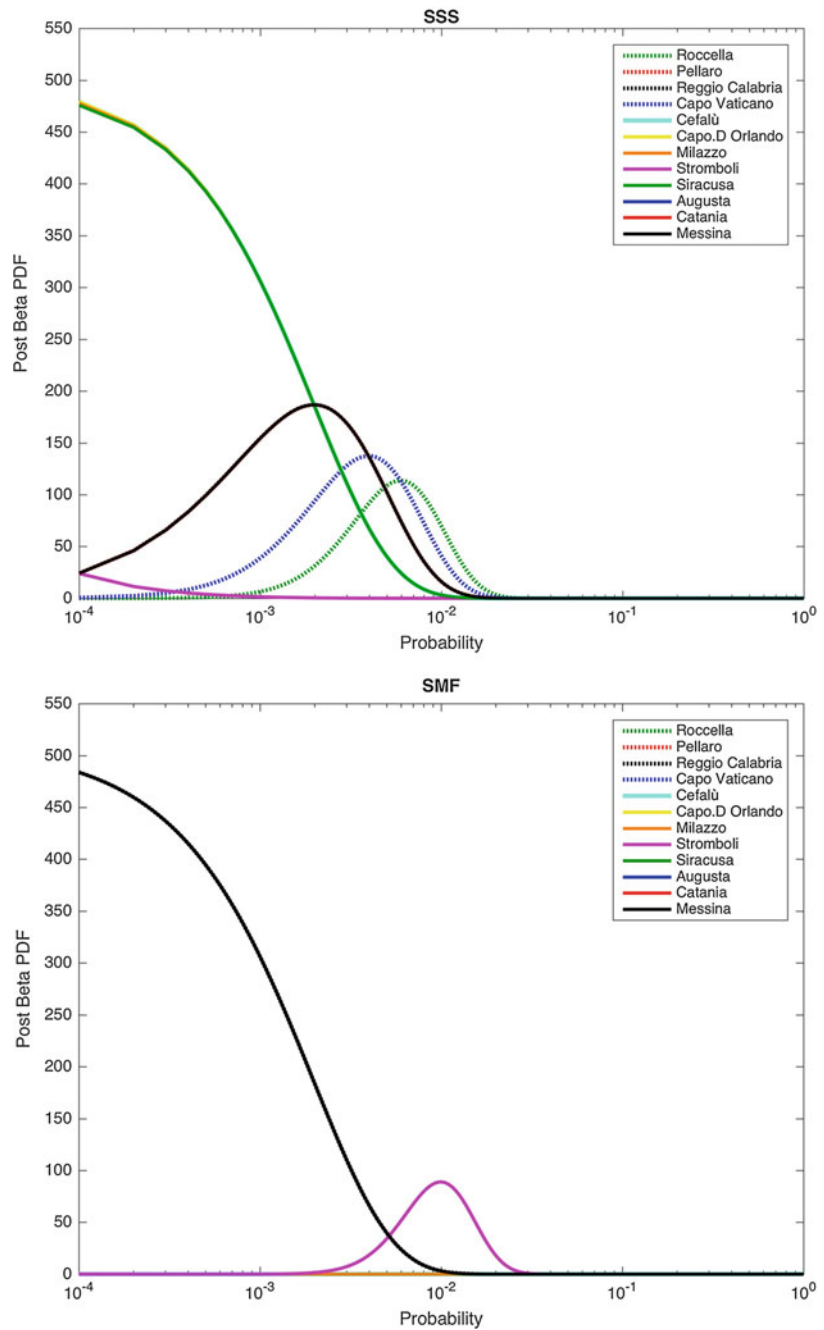


Tsunamis: Bayesian Probabilistic Analysis, Fig. 8 Prior *Beta* distributions for the (a) *SSS*s and (b) *SMF*s

The analysis shows that *SSS* and *SMF* posterior probability generally increases by one or more order of magnitude, and both types of tsunamigenic sources present the same order of magnitude in the Messina Strait Area. Therefore, both sources must be considered and combined in

Tsunamis: Bayesian Probabilistic Analysis,

Fig. 9 Posterior *Beta* distributions for the (a) *SSSs* and (b) *SMFs*



order to produce a reliable PTHA in this area. Conversely, the posterior variances are reduced by one order of magnitude in the *SSS* case and by two orders of magnitude in the *SMF* case. The epistemic uncertainty decreases when the number

of past data and/or historical information increase and the *Beta* distribution results more spiked because of $A_{\text{post}} = A + y^i$.

Finally, if Θ defines the probability of occurrence of at least one tsunami event in the time

interval, then $1 - \Theta$ is the generic probability that no tsunami occurs. The final posterior distribution is

$$P(\Theta)_{\text{post}} = 1 - [(1 - \Theta^{\text{SSS}}_{\text{post}})(1 - \Theta^{\text{SMF}}_{\text{post}})] \quad (14)$$

and evaluates the probability for the Sicily and Calabria cities that a tsunami run-up overcoming the threshold z_t occurs in the time interval of 1 year caused both by *SSSs* or *SMFs*. The mean and the variance of the final posterior distributions are reported in Table 2.

Discussion and Conclusions

In Bayesian method applied to tsunami hazard, the following issues should be taken into account and discussed:

- A mathematical consequence of the Bayesian procedure is that the results are always within the range of hypotheses, similar to most of the statistical techniques. Thus for the Bayesian methods to be considered objective, the results must depend on the assumed prior statistical model and observed data. In fact, different prior parameter determinations (with their own probability distributions considered as the random variables) may reach different

conclusions, in particular when few past data are available.

- Information at the base of prior PDFs and likelihood functions sometimes cannot be completely independent in practical applications, leading to potential double counting. For example, in long-term applications where all the data of rare events should be considered, this issue can seriously affect Bayesian inferences if not properly accounted for (e.g., discussion in Selva and Sandri 2013). To overcome this issue, in practice, two assertions can be considered: (i) the tsunami data used to define the prior probability models should be extracted by different catalogs (e.g., generic earthquakes, not necessarily tsunamigenic earthquakes) and/or used only in aggregated forms (e.g., the prior information may be derived by tsunami events that occurred globally and support general knowledge about tsunami processes), whereas (ii) the data used to create the likelihood functions should be mainly local, in order to reduce as much as possible the potential effect of possible double counting. Therefore all assumptions made in formulating the prior probabilities and in selecting the parameters should be stated explicitly so that the results can be properly assessed.

At the same time, Bayesian statistics presents several advantages in the present context. In fact:

Tsunamis: Bayesian Probabilistic Analysis, Table 2 Final means and variance of the posterior probability distribution that a tsunami run-up overcoming 0.5 m occurs in the time interval of 1 year due to the *SSSs* and *SMFs*

Key sites	Mean $\times 10^{-3}$	Variance $\times 10^{-5}$
Messina	5.9	1.1
Reggio Calabria	7.9	1.5
Pellaro	5.9	1.1
Catania	5.9	1.2
Augusta	5.9	1.2
Siracusa	3.9	0.8
Milazzo	2.0	0.4
Capo d'Orlando	2.0	0.4
Cefalù	2.0	0.4
Stromboli	11.9	2.3
Capo Vaticano	5.9	1.2
Roccella	7.9	1.6

- It enables merging of several kinds of available information in a homogeneous framework. Different statistical methods, theoretical deductions, background knowledge, physical beliefs, empirical laws, numerical models, analytical results, historical data, and instrumental measurements are combined and integrated. Bayesian techniques make use of all information, even sparse data, while keeping track of the assumptions about the prior knowledge or the level of ignorance. For a given probability model, an update of the final inferences is possible as soon as new models and/or additional data become available.
- It enables accounting for different sources of uncertainty, i.e., aleatory and epistemic uncertainty. The uncertainties are specified and synthesized in the statistical distributions, and the Bayesian procedure, considering potential all sources of information, enables a quantification and, in principle, a controlled reduction of the inherent epistemic uncertainties.
- It allows for propagating all the uncertainties from all the levels of the assessment. The most relevant sources of uncertainty from the tsunami source generation process to wave propagation and impact on the coasts may be reported and incorporated in the tsunami hazard computation (Marzocchi et al. 2004, 2008; Grezio et al. 2010, 2012; Gelman et al. 2013; Knighton and Bastidas 2015; Selva et al. 2016). Additionally, different types of potentially tsunamigenic sources may be included in the analysis in order to reduce biases (Grezio et al. 2015).

Future Directions

A key role in the future is PTHA testability against real and independent data. In its Bayesian interpretation, the probability represents a state of knowledge, and it is intrinsically subjective because all probabilities are degrees of belief that cannot be measured (Lindley 2000) and/or eventually rejected (Jaynes 2003). In PTHA, this means that the probabilistic quantification strictly refers to the next time window, and its results cannot be tested. The frequentist interpretation

instead intrinsically connects the probability definition to a measurable quantity (the past frequency) that can be theoretically known by analyzing an “infinite” sequence of outcomes for repeatable event (Popper 1983). This makes such frequencies formally testable against real data. An *unificationist* approach (Marzocchi and Jordan 2014) may then be adopted for PTHA, in which the expert opinion is regarded as a model distribution describing the long-run frequencies determined by the data-generating process. These frequencies, which characterize the aleatory variability, have epistemic uncertainty described by the experts’ distributions. As far as the knowledge of the system increases, our capability of assessing the true value of such frequencies is refined, that is, the epistemic uncertainty is reduced. Therefore, following this definition for the PTHA and related uncertainty, if “infinite” dataset is made available, Bayesian and classical PTHA will lead to equivalent results, since any subjective choice regarding priors is completely overcome by the infinite dataset perfectly constraining the long-run frequencies. However, we are unfortunately far from this case, being tsunamis relatively rare events as compared to our observation window.

Acknowledgments We wish to thank Gareth Davies and Eric Geist for the constructive comments during the review process.

Bibliography

- Aida I (1969) Numerical experiments for the tsunami propagation – the 1964 Niigata tsunami and the 1968 Tokachi-Oki tsunami. *Bull Earthquake Res Inst* 47:673–700
- Annaka T, Satake K, Sakakiyama T, Yanagisawa K, Shuto N (2007) Logic-tree approach for probabilistic tsunami hazard analysis and its applications to the Japanese coasts. *Pure Appl Geophys* 164:577–592
- Blaser L, Ohrnberger M, Riggelsen C, Babeyko A, Scherbaum F (2011) Bayesian network for tsunami early warning. *Geophys J Int* 185(3):1431–1443. <https://doi.org/10.1111/j.1365-246X.2011.05020.x>
- Box GEP, Tiao GC (1992) Bayesian inference in statistical analysis. Wiley Classics Library, New York, p 588
- Burbidge D, Cummins PR, Mleczko R, Thio HK (2008) A probabilistic tsunami hazard assessment for Western Australia. *Pure Appl Geophys*. <https://doi.org/10.1007/S00024-008-0421-X>

- Castello B, Olivieri M, Selvaggi G (2007) Local and duration magnitude determination for the Italian earthquake catalogue (1981–2002). *Bull Seismol Soc Am* 97:128–139
- Congdon P (2006) Bayesian statistical modelling, Wiley series in probability and statistics. Wiley, Chichester, p 529
- Davies G, Griffin J, Løvholt F, Glymsdal S, Harbitz C, Thio HK, Lorito S, Basili R, Selva J, Geist E, Baptista MA (2016) A global probabilistic tsunami hazard assessment from earthquake sources, Accepted Manuscript, “Tsunamis: geology, hazards and risks”. GSL Special Publications, London
- Draper D (2009) Bayesian statistics. *Enc Complexity Earth Syst Sci* 1:445–476
- Favalli M, Boschi E, Mazzarini F, Pareschi MT (2009) Seismic and landslide source of the 1908 Straits of Messina tsunami (Sicily, Italy). *Geophys Res Lett* 36:L16304. <https://doi.org/10.1029/2009GL039135>
- Geist EL (2002) Complex earthquake rupture and local tsunamis. *J Geophys Res* 107:ESE2–1–ESE 2–16
- Geist EL, Lynett PJ (2014) Source processes for the probabilistic assessment of tsunami hazards. *Oceanography* 27(2):86–93. <https://doi.org/10.5670/oceanog.2014.43>
- Geist EL, Oglesby DD (2014) Tsunamis: stochastic models of occurrence and generation mechanisms. In: Meyers RA (ed) *Encyclopedia of complexity and systems science*. Springer, New York. https://doi.org/10.1007/978-3-642-27737-5_595-1
- Geist EL, Parsons T (2006) Probabilistic analysis of tsunami hazards. *Nat Hazards* 37:277–314. <https://doi.org/10.1007/s11069-005-4646-z>
- Geist EL, Parsons T (2010) Estimating the empirical probability of submarine landslide occurrence. In: Mosher DC et al (eds) *Submarine mass movements and their consequences*. Advances in natural and technological hazards research, vol 28. Springer, Dordrecht, p 377
- Geist EL, Ten Brink US, Gove M (2014) A framework for the probabilistic analysis of meteotsunamis. *Nat Hazards* 74:123–142. <https://doi.org/10.1007/S11069-014-1294-1>
- Gelman A, Carlin JB, Stern HS, Rubin DB (2013) *Bayesian data analysis*. Chapman & Hall/CRC Press, Boca Raton, p 667
- Gonzalez FI, Geist EL, Jaffe B, Ka'noğlu U, Mofjeld H, Synolakis CE, Titov VV, Arcas D, Bellomo D, Carlton D, Horning T, Johnson J, Newman J, Parsons T, Peters R, Peterson C, Priest G, Venturato A, Weber J, Wong F, Yalciner A (2009) Probabilistic tsunami hazard assessment at Seaside, Oregon, for near- and far-field seismic sources. *J Geophys Res* 114:C11023. <https://doi.org/10.1029/2008JC005132>
- Gregory P (2005) *Bayesian logical data analysis for the physical sciences*. Cambridge University Press, Cambridge, p 468
- Grezio A, Marzocchi W, Sandri L, Gasparini P (2010) A Bayesian procedure for Probabilistic Tsunami Hazard Assessment. *Nat Hazards* 53:159–174. <https://doi.org/10.1007/s11069-009-9418-8>
- Grezio A, Marzocchi W, Sandri L, Argnani A, Gasparini P (2012) Probabilistic tsunami hazard assessment for messina strait area (Sicily – Italy). *Nat Hazards*. <https://doi.org/10.1007/s11069-012-0246-x>
- Grezio A, Tonini R, Sandri L, Pierdominici S, Selva J (2015) A methodology for a comprehensive probabilistic tsunami hazard assessment: multiple sources and short-term interactions. *J Mar Sci Eng* 3:23–51. <https://doi.org/10.3390/jmse3010023>
- Grezio A, Babeyko A, Baptista MA, Behrens J, Costa A, Davies G, Geist EL, Glimsdal S, González FI, Griffin J, Harbitz CB, LeVeque RJ, Lorito S, Løvholt F, Omira R, Mueller C, Paris R, Parsons T, Polet J, Power W, Selva J, Sørensen MB, Thio HK (2017) Probabilistic tsunami hazard analysis: multiple sources and global applications. *Rev Geophys* 55. <https://doi.org/10.1002/2017RG000579>
- Grilli ST, Watts P (2005) Tsunami generation by submarine mass failure, I: modeling, experimental validation, and sensitivity analyses. *J Waterway Port Coast Ocean Eng* 131:283–297
- Gutenberg B, Richter C (1944) Frequency of earthquakes in California. *Bull Seism Soc Am* 34:185–188
- Heidarzadeh M, Kijko A (2011) A probabilistic tsunami hazard assessment for the makran subduction zone at the Northwestern Indian Ocean. *Nat Haz* 56:577–593
- Hoechner A, Babeyko AY, Zamora N (2016) Probabilistic tsunami hazard assessment for the Makran region with focus on maximum magnitude assumption. *Nat Hazards Earth Syst Sci* 16:1339–1350. <https://doi.org/10.5194/nhess-16-1339-2016>
- Horspool N, Pranantyo I, Griffin J, Latief H, Natawidjaja DH, Kongko W, Cipta A, Bustaman B, Anugrah SD, Thio HK (2014) A probabilistic tsunami hazard assessment for Indonesia. *Nat Hazards Earth Syst Sci* 14:3105–3122. <https://doi.org/10.5194/nhess-14-3105-2014>
- Jaynes ET (2003) In: Bretthorst GL (ed) *Probability theory the logic of science*. Cambridge University Press, Cambridge, p 727
- Kagan YY (2002a) Seismic moment distribution revisited: I, statistical results. *Geophys J Int* 148:520–541
- Kagan YY (2002b) Seismic moment distribution revisited: II, moment conservation principle. *Geophys J Int* 149:731–754
- Kagan YY, Jackson DD (2000) Probabilistic forecasting of earthquakes. *Geophys J Int* 143:438–453
- Knighton J, Bastidas LA (2015) A proposed probabilistic seismic tsunami hazard analysis methodology. *Nat Hazards*. <https://doi.org/10.1007/s11069-015-1741-7>
- Lane EM, Gillibrand PA, Wang X, Power W (2013) A probabilistic tsunami hazard study of the auckland region, part II: inundation modelling and hazard assessment. *Pure Appl Geophys* 170:1635–1646. <https://doi.org/10.1007/s00024-012-0538-9>
- Lay T (2015) The surge of great earthquakes from 2004 to 2014. *Earth Planet Sci Lett Invited Front Pap* 409:133–146. <https://doi.org/10.1016/j.epsl.2014.10.047>
- Lindley DV (2000) The philosophy of statistics. *Statistician* 49:293–337
- Lorito S, Selva J, Basili R, Romano F, Tiberti MM, Piatanesi A (2015) Probabilistic hazard for seismically induced tsunamis: accuracy and feasibility of

- inundation maps. *Geophys J Int* 200(1):574–588. <https://doi.org/10.1093/gji/ggu408>
- Lorito S, Romano F, Lay T (2016) Tsunamigenic earthquakes (2004–2013): source processes from data inversion. In: Meyers RA (ed) *Encyclopedia of complexity and systems science*, vol 2015. Springer Science+Business Media, New York. https://doi.org/10.1007/978-3-642-27737-5_641-1
- Maramai A, Graziani L, Alessio G, Burrato P, Colini L, Cucci L, Nappi R, Nardi A, Vilardo G (2005a) Near- and far field survey report of the 30 December 2002 Stromboli (Southern Italy) tsunami. *Mar Geol* 215(93):106
- Maramai A, Graziani L, Tinti S (2005b) Tsunami in the Aeolian Islands (southern Italy): a review. *Mar Geol* 215(11):21
- Marzocchi W, Jordan TH (2014) Testing for ontological errors in probabilistic forecasting models of natural systems. *PNAS* 111(33):11973–11978. <https://doi.org/10.1073/pnas.1410183111/-DCSupplemental>
- Marzocchi W, Lombardi AM (2008) A double branching model for earthquake occurrence. *J Geophys Res* 113: B08317. <https://doi.org/10.1029/2007JB005472>
- Marzocchi W, Sandri L, Gasparini P, Newhall C, Boschi E (2004) Quantifying probabilities of volcanic events: the example of volcanic hazard at Mount Vesuvius. *J Geophys Res* 109:B11201. <https://doi.org/10.1029/2004JB003155>
- Marzocchi W, Sandri L, Selva J (2008) BET_EF: a probabilistic tool for long- and short-term eruption forecasting. *Bull Volcanol* 70:623–632. <https://doi.org/10.1007/s00445-007-0157-y>
- Marzocchi W, Taroni M, Selva J (2015) Accounting for epistemic uncertainty in PSHA: logic tree and ensemble modeling. *Bull Seismol Soc Am* 105:2151–2159. <https://doi.org/10.1785/0120140131>
- Mueller C, Power W, Fraser S, Wang X (2015) Effects of rupture complexity on local tsunami inundation: implications for probabilistic tsunami hazard assessment by example. *J Geophys Res Solid Earth* 120:488–502. <https://doi.org/10.1002/2014JB011301>
- O’Loughlin KF, Lander JF (2003) *Caribbean tsunamis: a 500-year history from 1498–1998*. Kluwer Academic Publishers, Dordrecht
- Okada Y (1985) Surface deformation due to shear and tensile faults in a half-space. *Bull Seismol Soc Am* 75:1135–1154
- Okada Y (1992) Internal deformation due to shear and tensile faults in a half-space. *Bull Seismol Soc Am* 82:1018–1040
- Omira R, Baptista MA, Matias L (2015) Probabilistic tsunami hazard in The Northeast Atlantic from near- and far-field tectonic sources. *Pure Appl Geophys* 172:901–920. 2014 Springer, Basel. <https://doi.org/10.1007/S00024-014-0949-X>
- Orfanogiannaki K, Papadopoulos G (2007) Conditional probability approach of the assessment of tsunami potential: application in three tsunamigenic regions of the Pacific Ocean. *Pure Appl Geophys* 164:593–603
- Parsons T (2008) Monte Carlo method for determining earthquake recurrence parameters from short paleoseismic catalogs: example calculations for California. *J Geophys Res* 113. <https://doi.org/10.1029/2007JB004998>
- Parsons T, Geist EL (2009) Tsunami probability in the Caribbean region. *Pure Appl Geophys* 165(2008):2089–2116. <https://doi.org/10.1007/s00024-008-0416-7>
- Paté-Cornell M (1996) Uncertainties in risk analysis: six levels of treatment. *Reliab Eng Syst Saf* 54:95–111
- Polet J, Kanamori H (2009) Tsunami earthquakes. In: Meyers A (ed) *Encyclopedia of complexity and systems science*. Springer, New York. https://doi.org/10.1007/978-0-387-30440-3_567
- Popper KR (1983) *Realism and the aim of science*. Hutchinson, London
- Power W, Wang X, Lane EM, Gillibrand PA (2013) A probabilistic tsunami hazard study of the Auckland region, part I: propagation modelling and tsunami hazard assessment at the shoreline. *Pure Appl Geophys* 170:1621. <https://doi.org/10.1007/s00024-012-0543-z>
- Reid RO, Bodine BR (1968) Numerical model for storm surges in Galveston Bay. *J Waterways and Harbors Div ACE* 94:33–57
- Sakai T, Takeda T, Soraoka H, Yanagisawa K, Annaka T (2006) development of a probabilistic tsunami hazard analysis in Japan. In: *Proceedings of ICONE14 international conference on nuclear engineering*, Miami, 17–20 July, ICONE14–89183
- Satake K (1995) Linear and nonlinear computations of the 1992 Nicaragua earthquake tsunami. *Pure Appl Geophys* 144:455–470
- Satake K (2002) Tsunamis. In: Lee WHK, Kanamori H, Jennings PC, Kisslinger C (eds) *International handbook of earthquake and engineering seismology*, vol 81A. Academic Press, Amsterdam, pp 437–451
- Selva J, Sandri L (2013) Probabilistic seismic hazard assessment: combining cornell-like approaches and data at sites through Bayesian inference. *Bull Seismol Soc Am* 103(3):1709–1722. <https://doi.org/10.1785/0120120091>
- Selva J, Tonini R, Molinari I, Tiberti MM, Romano F, Grezio A, Melini D, Piatanesi A, Basili R, Lorito S (2016) Quantification of source uncertainties in seismic probabilistic tsunami hazard analysis (SPTHA). *Geophys J Int* 2016. <https://doi.org/10.1093/gji/ggw107>
- Shin JY, Chen S, Kim T-W (2015) Application of Bayesian Markov Chain Monte Carlo Method with mixed gumbel distribution to estimate extreme magnitude of tsunamigenic earthquake. *KSCE J Civ Eng* 19(2):366–375. <https://doi.org/10.1007/s12205-015-0430-0>
- Shuto N (1991) Numerical simulation of tsunamis – its present and near future. *Nat Hazards* 4:171–191
- Smith WHF, Sandwell DT (1997) Global seafloor topography from satellite altimetry and ship depth soundings. *Science* 277:1957–1962
- Sørensen MB, Spada M, Babeyko A, Wiemer S, Grünthal G (2012) Probabilistic tsunami hazard in the Mediterranean Sea. *J Geophys Res* 117:B01305. <https://doi.org/10.1029/2010JB008169>

- Stigler SM (1983) Who discovered Bayes Theorem? *Am Stat* 37:290–296
- Suppasri A, Imamura F, Koshimura S (2012) Probabilistic tsunami hazard analysis and risk to coastal populations in Thailand. *J Earthquake and Tsunami* 06:1250011. [27 Pages]. <https://doi.org/10.1142/S179343111250011x>
- Synolakis CE (1987) The runup of solitary waves. *J Fluid Mech* 185:523–545
- Tanioka Y, Satake K (1996) Tsunami generation by horizontal displacement of ocean bottom. *Geophys Res Lett* 23:861–865
- Tatsumi D, Calder CA, Tomita T (2014) Bayesian near-field tsunami forecasting with uncertainty estimates. *J Geophys Res Oceans* 119:2201–2211. <https://doi.org/10.1002/2013JC009334>
- Thio HK, Li W (2015) Probabilistic tsunami hazard analysis of the cascadia subduction zone and the role of epistemic uncertainties and aleatory variability. In: 11th Canadian conference on earthquake engineering, Victoria, pp 21–24
- Thio HK, Somerville P, Polet J (2010) Probabilistic tsunami hazard in California, PEER Report 2010/108 Pacific Earthquake Engineering Research Center
- Tinti S, Maramai A, Graziani L (2004) The new catalogue of Italian Tsunamis. *Nat Haz* 33(439):465
- Tinti S, Armigliato A, Tonini R, Maramai A, Graziani L (2005) Assessing the hazard related to tsunamis of tectonic origin: a hybrid statistical-deterministic method applied to Southern Italy coasts. *ISET J Earthquake Tech* 42:189–201
- Tinti S, Argnani A, Zaniboni F, Pagnoni G, Armigliato A (2007) Tsunamigenic potential of recently mapped submarine mass movements offshore eastern Sicily (Italy): numerical simulations and implications for the 1693 tsunami. IASPEI—JSS002—abstract n. 8235 IUGG XXIV General Assembly, Perugia, 2–13 July 2007
- TPSWG Tsunami Pilot Study Working Group (2006) Sea-side, Oregon tsunami pilot study—modernization of FEMA flood hazard maps. NOAA OAR Special Report, NOAA/OAR/PMEL, Seattle, p 94 + 7 appendices
- Vannucci G, Gasperini P (2004) The new release of the database of earthquake mechanisms of the mediterranean area (EMMA2). *Ann Geophys Suppl* 47:307–334
- Watts P, Grilli ST, Tappin D, Fryer GJ (2005) Tsunami generation by submarine mass failure, II: predictive equations and case studies. *J Waterway Port Coast Ocean Eng* 131:298–310
- Wells DL, Coppersmith KJ (1994) New empirical relationships among magnitude, rupture length, rupture width, rupture area and surface displacement. *Bull Seismol Soc Am* 84:974–1002
- Yadav RBS, Tsapanos TM, Tripathi JN, Chopra S (2013) An evaluation of tsunami hazard using Bayesian approach in the Indian Ocean. *Tectonophysics* 593:172–182

Tsunami Inundation, Modeling of

Patrick J. Lynett
Texas A&M University, College Station, TX,
USA

Article Outline

Glossary
Definition of the Subject
Introduction
Brief Review of Tsunami Generation and Open
Ocean Propagation
Physics of Nearshore Tsunami Evolution
Effects of Bathymetric and Topographical
Features on Inundation
Hydrodynamic Modeling of Tsunami Evolution
Moving Shoreline Algorithms
Future Directions
Bibliography

Glossary

Beach profile A cross-shore, or normal to the beach, survey of the seafloor and dry ground elevation (bathymetry and topography); a series of spatial location and bottom elevation data pairs.

Bore A steep hydraulic front which transitions between areas of different water level. Tsunamis can approach land as a turbulent, breaking bore if the incident tsunami is of sufficiently large height.

Boussinesq equations An approximate equation model, used for waves with wave length of at least two times the local water depth; a long-wave-based model, but includes some frequency dispersion

Dispersion, amplitude The separation of wave components due to a wave-height related difference in wave speed; all else being equal, a

wave with a large height will travel faster than one with a small height.

Dispersion, frequency The separation of wave components due to a frequency related difference in wave speed; all else being equal, a wave with a longer period will travel faster than one with a short period.

Navier–Stokes equations The full equations of fluid motion, including dissipation through the fluid molecular viscosity only. Other models discussed here, namely the Shallow Water Wave and Boussinesq equations, are approximations to these equations.

Runup, or runup height The ground elevation (a vertical measure) at the furthest point of landward inundation.

Shallow water wave equations An approximate equation model, used for waves with wave length many times larger than the water depth; a non-dispersive, long-wave model; there is no frequency dispersion in this model.

Tsunami inundation The spatial area flooded as a tsunami rushes inland.

Definition of the Subject

Tsunami inundation is the one of the final stages of tsunami evolution, when the wave encroaches upon and floods dry land. It is during this stage that a tsunami takes the vast majority of its victims. Depending on the properties of the tsunami (e.g. wave height and period) and the beach profile (e.g. beach slope, roughness), the tsunami may approach as a relatively calm, gradual rise of the ocean surface or as an extremely turbulent and powerful bore – a wall of white water. The characteristics of this approach determine the magnitude and type of damage to coastal infrastructure and, more importantly, the actions required of coastal residents to find a safe retreat or shelter.

To gage the nearshore impact of tsunami inundation, engineers and scientists rely primarily on three different methods: (1) Field survey of past

events, (2) Physical experimentation in a laboratory, and (3) Numerical modeling. It is the last of these methods – numerical simulation of tsunami inundation – that will be the focus of this article. With numerical simulation, it is possible to predict the consequence of future tsunamis on existing coastal towns and cities. This information allows for the establishment of optimum evacuation routes, identification of high-risk and/or unprepared areas, re-assessment of building codes to withstand wave impact, and placement of tsunami shelters, for example. It is the hope that, through accurate prediction of tsunami effects, in conjunction with policy makers willing to implement recommended changes and a strong public education program, communities will show resiliency to tsunami impact, with minimal loss of life and damage to critical infrastructure.

Introduction

On December 26, 2004, the boundary between the Indo–Australian and Eurasian plates off the coast of northern Sumatra ruptured in a great (Mw 9.3) earthquake at 00:58:53 universal time (U.T.). Up to 15 m of thrust on the plate interface (Lay et al. 2005) displaced tens of cubic kilometers of seawater and propagated a tsunami across the Indian Ocean. The earthquake was widely felt throughout South Asia and was locally destructive in Sumatra and the Andaman and Nicobar islands, but it was the tsunami that caused widespread damage to densely populated coastal communities both nearby and thousands of kilometers away.

Due to the extensive damage left behind by large tsunamis such as the Indian Ocean tsunami, it is difficult if not impossible to put together a complete picture of the event with field observations alone. Additionally, for some parts of the world that have not seen a tsunami in recent times, there are no field observations on which to develop safety procedures and protect residences from future tsunamis. It is for these purposes – understanding the detail of tsunami inundation and to estimate tsunami hazard – that we must rely on modeling of tsunamis. There are two primary modeling approaches – physical and

numerical. The physical, or experimental, approach uses scaled-down models to look at a particular aspect of a phenomenon. While this approach is integral to the fundamental understanding of waves, because of the huge wavelengths of tsunamis, experiments are limited. For example, a tsunami might have a wavelength of 100 km in a deep ocean depth of 1 km, with a wave height of 1 m. Note that the above values represent approximate order of magnitudes for a large subduction zone tsunami. Now, to scale this down for the laboratory with a wave tank depth of 1 m – the tank would have to be 100 m long and the created lab-tsunami would have a hardly measurable height of 1 mm. Numerical modeling, while not “real” in the sense that modeling is done on a computer chip with approximated equations of motion rather than in the laboratory, does not suffer from this scaling problem, and can generally accommodate any type of arbitrary wave and ocean depth profile.

Numerical simulations of tsunami propagation have been greatly improved in the last 30 years. In the United States, several computational models are being used in the National Tsunami Hazard Mitigation Program, sponsored by the National Oceanic and Atmospheric Administration (NOAA), to produce tsunami inundation maps and predict tsunami runup in real-time for the warning system. In addition, there are numerous other models used by researchers and engineering companies in an attempt to better understand tsunami impact. In this article, an overview of these models, as well as how they are validated and utilized, is provided.

Brief Review of Tsunami Generation and Open Ocean Propagation

Before introducing the physics behind propagating a tsunami across oceans and overland, we must first discuss how a tsunami is created. For earthquake generated mega-tsunamis, such as the Indian Ocean event, a huge undersea earthquake along a great fault length of a subduction zone must occur. These earthquakes create large vertical motions of the seafloor. This vertical motion of

the seafloor pushes the water above it, essentially creating a small displacement of water above the earthquake. This displacement of water will immediately try to spread out and reach a gravitational equilibrium, and it does so as waves propagating away from the earthquake zone – this is the tsunami.

To represent the tsunami in numerical models, we use an initial condition. Simply put, there is placed some irregular ocean surface profile at the instant after the earthquake, when the numerical simulations will start. Then, based on physics – Newton's Laws written for fluid – the initial condition evolves and transits oceans. As a tsunami travels unhindered across ocean basins, it does so quickly and with little noticeable change. In the deep ocean, even the largest tsunamis have heights only near 1 m and currents of 10 cm/s, and are not likely to be identified by ships or surface buoys in the presence of wind waves.

Physics of Nearshore Tsunami Evolution

A tsunami in the deep ocean is long and travels extremely fast. As the wave reaches shallow water, near the coastline, the tsunami begins the shoaling process. The speed at which long wave such as a tsunami moves, or celerity, is a function of the local water depth. The less the depth, the slower the wave moves. A tsunami, with its very long length, experiences different water depths at any given instant as it travels up a slope; the depth at the front of the wave, the portion of the tsunami closest to the shoreline, will generally be in the shallowest water and thus is moving the slowest. The back of the tsunami, on the other hand, will be in deeper water and will be moving faster than the front. This leads to a situation where the back part of the wave is moving faster than the front, causing the wavelength to shorten. With a shortening tsunami, the wave energy is in essence squeezed into a smaller region, forcing the height to grow. It is for this reason that, despite having a height of only a meter in the deep ocean, the tsunami elevation over land can easily exceed 10 m. With this great increase in wave height comes a more dynamic and complex phenomenon.

Presented in a more technical manner, a tsunami in the open ocean is generally a linear, non-dispersive wave. First, what is meant by a non-dispersive tsunami will be discussed. Also, the discussion here will be in terms of a large earthquake generation tsunami, such as the 2004 Indian Ocean event. Other impulsive waves, such as landslide or asteroid impact generated waves, are more difficult to generalize and will be introduced separately at the end of this section.

Any wave condition, whether it is a tsunami or a typical wind wave in the ocean, can be mathematically described as a superposition, or summation, of a series of separate sine (or cosine) waves, each with independent amplitude and speed. For example, with the right choice of individual sine waves, it is possible to construct even the idealized tsunami: a single soliton. If a wave is considered a dispersive wave, then the various sine wave components will have different wave speeds, and the wave will disperse as the faster moving components move away from the slower ones. If a wave is non-dispersive, then all the components move at the same speed, and there is no length-wise dispersal, or spreading, of the tsunami wave energy. It is for this reason that tsunamis can be devastating across such a large spatial region; the tsunami wave energy will not disperse but will remain in a focused pulse.

The dispersion described above is generally what scientists are referring to during a discussion of dispersive vs non-dispersive waves. However, it is more precisely called “frequency dispersion” as it is dominantly dependent on the period of the component. There is another type of dispersion, called “amplitude dispersion.” This second type of dispersion is a function of the nonlinearity of the wave, and is usually discussed under the framework of linear vs nonlinear waves. For tsunamis, the nonlinearity of the waves is given by the ratio of the tsunami height to the water depth. When this ratio is small, such as in the open ocean, the wave is linear; on the other hand, in shallow water the ratio is order unity and thus the wave is no longer linear. The linear/nonlinear nomenclature is not an intuitive physical description of the waves, but comes from the equations describing the tsunami motion, described later in this section.

When this nonlinear effect is taken into account, it is found that the wave speed is no longer just a function of the local depth, but of the wave height as well. More specifically, looking at two components of the same period but with different amplitudes, the component with the larger amplitude will have a slightly larger wave speed. Except for the interesting cases of wave fission, discussed later in this section, the nonlinear effect of amplitude dispersion does not spread tsunami energy with an end result of lessening nearshore impact; in fact it will act to focus wave energy at the front, often leading to a powerful breaking bore.

Thus, open ocean propagation of a conventional tsunami is a relatively uncomplicated process which translates wave energy across basins, subject to wave speed changes that are a function of the local depth. As a tsunami enters the nearshore region, roughly characterized by water depths of 100 m and less, the wave can undergo a major physical transformation. The properties of this transformation depend heavily on the characteristics of the beach profile and the wave itself. In the simplest inundation case, the beach profile is relatively steep (footnote: here “steep” should be thought of in terms of the tsunami wavelength. If the horizontal distance along the slope connecting deep water to the shoreline is small compared to the tsunami wavelength, the beach would be considered steep) and the tsunami wave height is small, then the runup process closely resembles that of a wave hitting a vertical wall, and the runup height will be approximately twice the offshore tsunami height. In these special cases, a breaking bore front would not be expected; in fact horizontal fluid velocities near the shoreline would be very small. Here, the tsunami inundation would closely resemble that of a quickly rising tide with only very minor turbulent, dynamic impacts. However, even in these cases, overland flow constrictions and other features can create localized energetic inundation.

If the beach profile slope is mild, typical of continental margins, and/or the tsunami wave height is large, then the shallow water evolution process becomes highly nonlinear. However, while the nonlinear effect becomes very important, in the large majority of cases, frequency

dispersion is still very small and can be neglected. Nearshore nonlinear evolution is characterized by a strong steepening, and possible breaking, of the wave front with associated large horizontal velocities. In these cases, turbulent dissipation can play a major role.

While it may be intuitive to postulate that wave breaking dissipation at the tsunami front plays a significant role in the tsunami inundation, this may not be altogether correct. This breaking dissipation, while extraordinarily intense, is fairly localized at the front which, both spatial and temporal, often represents only a small fraction of the tsunami. So, for tsunamis such as the 2004 Indian Ocean event, the related dissipation likely had only minor impact on leading-importance quantities such as the maximum runup and inland (off-beachfront) flow velocities. However, the properties of breaking are of great importance to other aspects of tsunami inundation. The maximum forces on beachfront infrastructure, such as ports, terminals, piers, boardwalks, and houses, should include the bore impact force as well as the drag force associated with the following quasi-steady flow (Ramsden 1996; Yeh 2006). If one was interested in understanding how bottom sediments are suspended, transported, and deposited by a tsunami, the bore turbulence again may play an important role. Thus, understanding the dynamics of a breaking tsunami front is not of particular importance for near real-time or operational tsunami forecast models. This information is of great use for engineers and planners, who can utilize it to design tsunami-resistant structures, for example.

A second energy dissipation mechanism, one that does play a major role in determining maximum runup, is bottom friction. On a fundamental level, this dissipation is caused by the flow interaction with the bottom, where bottom irregularities lead to flow separations and the resulting turbulence. All natural bottoms result in some bottom friction; a smooth, sandy beach may generate only minor dissipation, while a coral reef or a mangrove forest can play a huge role in reducing tsunami energy (Fernando et al. 2005). Such features will be discussed in additional detail in the next section. Other means of energy dissipation

will be largely local, and may include enhanced mixing due to sediment or debris entrainment, large shallow-flow vortex generation by headlands or other natural or artificial obstacles and the resulting dissipation, and flow through/around buildings and other infrastructure, sometimes termed macro-roughness and grouped with bottom friction.

Up to this point, we have only discussed the “typical” nearshore tsunami evolution which is portrayed as a wave without frequency dispersion, and may be called a linear or nonlinear tsunami, depending on a number of physical properties. The rest of this section will be devoted to those situations where the above characterization may no longer be adequate. Looking first to the tsunami source, waves that are generated by underwater landslides, underwater explosions, or asteroid impacts will often not behave as non-dispersive waves in the open ocean (Lynett et al. 2003; Weiss et al. 2006). These source regions tend to be at least an order of magnitude smaller in spatial extent compared to their subduction zone counterparts. Physically, this implies that the generated waves will be of shorter wavelength. As a rule of thumb, if these generated waves have length scales of less than 10 times the local depth, then it should be anticipated that frequency dispersion will play a role (Lynett and Liu 2002). Under this constraint, individual component wave speeds near the dominant period become frequency dependent.

Understanding that an impulsively generated wave can be dispersive has serious implications. Take, for example, a hypothetical landslide located in the Atlantic Ocean which generates a dispersive tsunami (e.g. Ward and Day 2001). As this tsunami travels across the Atlantic, to either the USA east coast or the European west coast, frequency dispersion effects will spread the wave energy in the direction of propagation. This will convert the initial short-period pulse into a long train of waves. By spreading this energy out, the inundation impact will be greatly reduced. First, by taking a high-density energy pulse and stretching it into a longer, lower-density train, the maximum energy flux, and thus intensity, hitting the shoreline will decrease. Second, by

increasing the duration of the time series, and creating many individual crests, energy dissipation can play a bigger role. Using simple energy arguments, it can be shown that, comparing a high-density, short-period pulse to a low-density, long-period train, more energy will be removed through bottom friction and breaking. This increase will be related to the ratio of the period of the entire dispersive wave train to the period of the pulse. Numerically studies have shown that for such cases, the individual wave crests are largely dissipated, and runup is dominated by the carrier wave, or in other terms it becomes a time-dependent, wave setup problem (Korycansky and Lynett 2007).

While a topic of current research in the tsunami community, frequency dispersion may occasionally play a non-negligible role in even the long wavelength, subduction zone tsunamis. To date, there have been two categories of argument that dispersion is important for these tsunamis: (1) short-period energy generated at the source is significant and leads to different patterns of runup if included (e.g. Horrillo et al. 2006; Kulikov 2005), and (2) shallow-water nonlinear interactions can generate short-period components which can become decoupled (or un-locked) from the primary wave, and will change the incident tsunami properties (Matsuyama et al. 2007).

Thinking of an arbitrary and complex initial free surface displacement generated by an under-sea earthquake, there does exist the possibility that dispersive wave energy can be initially generated here. This irregular wave condition can be constructed as a continuous wave energy spectrum, and by definition there will be finite (albeit small) energy at all frequencies. The obvious question in this case is: what length scale characteristics of the initial free surface displacement, or the preceding earthquake, will lead to a significant measure of dispersive wave energy? To provide an answer to this question, a simple order-of-magnitude scaling argument is presented here; see Hammack and Segur (1978), for example, for a mathematically rigorous attempt at insight. Let us define a characteristic change in vertical free surface elevation, $\Delta\eta$, and a horizontal length scale across which this vertical change occurs, ΔL .

Reducing to the simplest case, a regular wave with wave height equal to $\Delta\eta$, then the wavelength would be $2\Delta L$. Following this analogy, which will hold in a proportional sense for a Fourier series of wave components, for any $\Delta\eta$ measured along a tsunami initial condition, there exists a wave component with wave length equal to $2\Delta L$. For that component to be significant to the tsunami evolution, the local vertical change, $\Delta\eta$, must be some non-negligible fraction of the maximum tsunami height, H . Note that the H discussed here is a global property of the entire initial tsunami wave condition. For the individual wave component, with length $2\Delta L$, to be dispersive, its length should be less than roughly 10 times the local water depth, h_0 . Additionally, the difference between dispersive wave propagation and a non-dispersive propagation is cumulative. For example, if the full linear wave theory predicts, or a specific wave component, a wave speed that is 10% less than the long wave speed, then the predicted arrival time difference will grow by 0.1 times the period for each wavelength of propagation. Thus, the impact of dispersion is related to the distance of propagation, and is proportional to D/λ , where D is the total distance traveled by the wave, and λ is the average wavelength of the wave across D which can be expected to be proportional to $2\Delta L$. Assuming that λ is approximately equal to $2\Delta L$, it can be said that in order for frequency dispersion effects to play a role in tsunami evolution,

$$\max \left| \frac{h_0 \Delta\eta}{\Delta L^2} \right| \frac{D}{H} > \delta, \quad (1)$$

where δ is some minimum threshold for importance. What this value should be is an open question, although it is likely to be near 0.1. From this term, it is clear the impact of dispersion is a function of a number of the properties of the initial tsunami condition, and should be taken into consideration when creating tsunami initial conditions. For example, use of discontinuous block-type segments (e.g. Ioualalen et al. 2007), with sharp edges (very small ΔL) may lead to the conclusion that frequency dispersion is important, while it could be a direct result of a coarsely

approximated initial tsunami condition. Also note that this exercise does not include the effects of radial spreading, which could very likely be important for small-scale irregularities in the initial condition. Wave height decrease by radial spreading is proportional the horizontal curvature of the initial condition and to $(\lambda/D)^n$, and decreases faster for dispersive waves ($n \sim 1$) as compared to non-dispersive waves ($n = 0.5$) (e.g. Weiss et al. 2006). Thus, for this case of significant radial spreading, it would be very difficult for source-based dispersion effects to play a meaningful role in the far field.

Under certain conditions, namely a nonlinear tsunami propagating across a wide shallow shelf, a process called fission may occur. Wave fission is a separation process where wave energy, initially part of a primary wave or pulse, attains certain properties, such as higher or lower phase speed, that allow it to disconnect from the primary wave and propagate as an independent wave. In the context of nearshore tsunami evolution, there is a standard mechanism which is the cause of this fission. First, it is necessary to describe what a nonlinear, phase-locked wave is. To do this, we will examine the acceleration terms of the 1D conservation of momentum equation for the velocity component u :

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} = -\frac{\partial p}{\partial x} + \mu \frac{\partial^2 u}{\partial x^2}. \quad (2)$$

Now assume that there is a single wave component, under which the velocity oscillates as

$$\cos(kx - \omega t) = \cos \theta, \quad (3)$$

where k is the wavenumber, ω the frequency, and the speed of the wave is given by ω/k . If the wave is nonlinear, which is to say that the convective acceleration term in the above momentum equation is not negligible, the convective term will include the product of

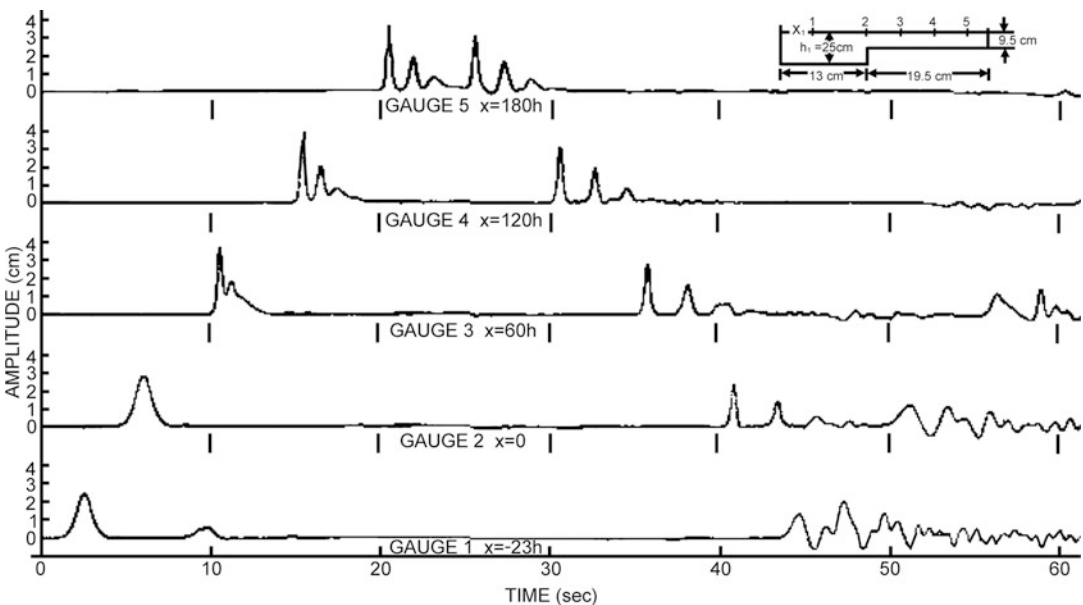
$$\begin{aligned} \cos \theta * \cos \theta &= \cos 2\theta \\ &= \cos(2kx - 2\omega t). \end{aligned} \quad (4)$$

Thus, through this nonlinear term, a new wave component, with twice the wavenumber and

frequency (or half the wavelength and period) has been generated. From linear wave theory, it is expected that this new wave, with a shorter period, will have a different wave speed than the original, primary wave. However, from the phase function of this new wave, there is the speed $2\omega/2k = \omega/k$, which is the identical speed of the primary wave. Thus this new wave is *locked* to the *phase* of the original wave. This connection can be rather delicate, and any disruptions to the primary wave, such as a varying seafloor, dissipation, or interactions with other free waves in the train or wave pulse, can cause the new waves to become unlocked. When this occurs, the now free waves retain their frequency 2ω , but take a wavenumber as given by linear wave theory. Since these freed waves will be of a shorter period than the primary wave, they will travel at a slower speed and generally trail the main wave front.

Long wave fission is most commonly discussed in the literature via a solitary wave propagating over an abrupt change in depth, such as a step (e.g. Goring and Raichlen 1992; Johson 1972; Losada et al. 1989; Liu and Cheng

2001; Madsen and Mei 1969; Seabra-Santos et al. 1987). In these cases, there is a deep water segment of the seafloor profile, where a solitary wave initially exists. In this depth, the solitary wave is of permanent form. As the solitary wave passes over the change in depth, into shallower water, the leading wave energy will try to re-discover a balance between nonlinearity and dispersion; the solitary wave. Since this new solitary wave will be a different shape and contain a lower level of mass, by conservation there must be some trailing disturbance to account for the deficient. This trailing disturbance will take the form of a rank-ordered train of solitons. Figure 1 depicts this process. The solitons in the trailing train, while smaller in height than the leading solitary wave, tend to have a similar wavelength; this has been shown both analytically, numerically, and experimentally. Note, however, that discussion of fission in this sense is not particularly relevant to “real” tsunami modeling, where the offshore wave approaching the shelf break rarely resembles a solitary wave solution (Tadepalli and Synolakis 1996). However, the offshore wave does not need



Tsunami Inundation, Modeling of, Fig. 1 Example of experimental data looking at solitary wave fission by propagation onto a shelf from Goring and Raichlen (1992). Note that the flume layout and measurement location is

given up in the upper right. The initial solitary wave undergoes the fission process and results in three distinct solitary waves

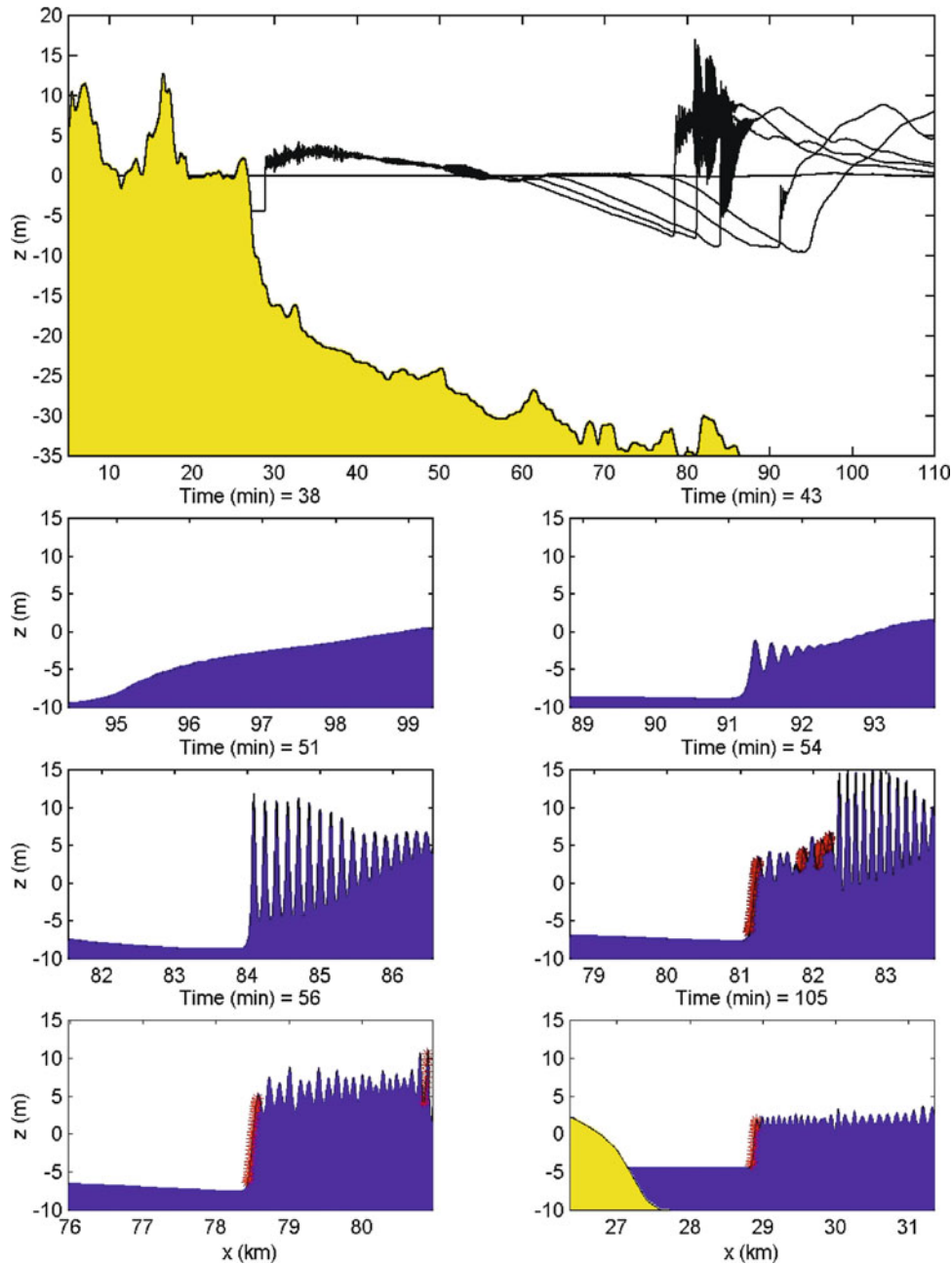
to specifically be a solitary wave for this process to occur.

In numerous eyewitness accounts and videos recorded of the 2004 Indian Ocean tsunami, there is evidence of the tsunami approaching the coastline as a series of short period (on the order of 1 min and less) breaking fronts, or bores (e.g. Ioualalen et al. 2007). These short period waves may be the result of fission processes of a steep tsunami front propagating across a wide shelf of shallow depth. Along the steep front of a very long period wave, nonlinearity will be very important. There will be a large amount of energy in high-frequency components with wavelengths similar the horizontal length of the tsunami front (on the order of 1 km). As the wave continues to shoal, the high-frequency locked waves may eventually become free waves, and will take the form of very short waves “riding” the main wave pulse. This situation is akin to an undular bore in a moving reference frame. This process is, in fact, identical to that described in the above paragraph, it simply takes place over a much longer distance. The newly freed waves, in the nonlinear and shallow environment, will attempt to reach an equilibrium state, where frequency dispersion and nonlinearity are balanced. Thus, the fission waves will appear as solitary waves, or more generally, cnoidal waves. This fact provides some guidance as to the wavelength of these fission waves; they can be approximately calculated via solitary wave theory using the tsunami height and depth of the shelf. For example, on a shelf with depth of 30 m and an incident tsunami height of 5 m, fission waves with a wavelength of approximately 240 m and period of 13 s would be generated. In recent work looking at tsunamis along the eastern USA coast, where there exists a wide shallow shelf, this fission process has been investigated (Geist et al. 2008). Figure 2 gives a few numerical simulation snapshots, and shows where the fission occurs, and the eventually impact on the waveform. This simulation, run with the dispersive equations, generated fission waves with lengths in the range of 100–200 m, and required a grid size of 5 m to attain numerically convergent results. In this

example, the steep fission waves break offshore, and have little impact on the maximum runup. A conclusion of this fission issue is that, if one attempts to simulate tsunami propagation with dispersive equations, and if the grid is not chosen to be fine enough to resolve the short fission waves, the justification to use the dispersive model is greatly degraded.

Effects of Bathymetric and Topographical Features on Inundation

It is well established that large-scale coastal features, such as small islands, large shoals, canyons, and shelves, can play an important role in tsunami inundation due to conventional shallow water effects such as shoaling and refraction (e.g. Briggs et al. 1994; Carrier 1966; González et al. 1995; Liu et al. 1995; Yeh et al. 1994). On the other hand, understanding of the impact of smaller scale features is just now being developed. This work was largely initiated by field observations. Synolakis et al. (1995), surveying the coast of Nicaragua for information about the 1992 tsunami in the region, noted that the highest levels of damage along a particular stretch of beach were located directly landward of a reef opening used for boat traffic. It was postulated that the reef gap acted as a lower resistance conduit for tsunami energy, behaving like a funnel and focusing the tsunami. Along neighboring beaches with intact reefs, the tsunami did not have the intensity to remove even beach umbrellas. Investigating impacts from the same tsunami, Borrero et al. (1997), discussed how small scale bathymetry variations affected coastal inundation. One of the conclusions of this work was that bathymetry features with length scales 50 m and less had leading order impact on the runup. Looking to the recent Indian Ocean tsunami, a survey team in Sri Lanka inferred from observations that reef and dune breaks lead to locally increased tsunami impact (Liu et al. 2005a, b). Also in Sri Lanka, Fernando et al. (2005) performed a more thorough survey along the southeastern coastline, and concluded that there was a compelling correlation between coral mining and locally severe tsunami



Tsunami Inundation, Modeling of, Fig. 2 Example of tsunami fission. Simulation results are from Geist et al. (2008) for a landslide-generated tsunami off the east coast of the USA. The top plot shows the beach profile and six free surface profiles at different times. The lower

subplots are zoom-in's of those six profiles, with the times given in the individual plot titles. The red marks visible in the lowest plots indicate regions where the wave is breaking

damage. While additional research is needed to quantify the effects of small scale features, the observations hint that defense measures such as

seawalls, once thought to be inconsequential to tsunami inundation, may provide some protection.

Onshore, tsunami propagation is effected by the general topography (ground slope), ground roughness, and obstacles (e.g. Lynett 2007; Synolakis 1987; Tadeipalli and Synolakis 1994; Tomita and Honda 2007). The composition of the ground, be it sand, grass, mangroves, or pavement, controls the roughness and the subsequent bottom friction damping. To predict tsunami inundation with high confidence, the ground type must be well mapped and the hydrodynamic interaction with that type must be well understood. If the tsunami approaches the shoreline as a bore, the process of “bore collapse”, or the conversion of potential to kinetic energy, will cause the fluid to rapidly accelerate (Shen and Meyer 1963; Yeh et al. 1989). This fast flow equates to high fluid forces on obstacles such as buildings. Tsunami interaction with these obstacles can lead to a highly variable local flow pattern (e.g. Cross 1967; Tomita and Honda 2007). As the flow accelerates around the corners of a building, for example, the scour potential of that flow increases greatly, and foundation undermining is a concern (e.g. see Fig. 3). As with any fluid flow past an obstacle, the backface of the obstacle is characterized by a low-pressure wake. Combined with the interior flooding of a building, this low pressure wake may lead to an outward “pull” force on the back wall, causing it to fail by falling away from the center of the building. Such failures were observed during field surveys of the 2004 event, as shown in Fig. 4. Increasing the topographical complexity, in built coastal environments, structures are located within close enough proximity to each other such that their disturbances to the flow may interact. This can lead to irregular and unexpected loadings, where for example a 2nd row building experiences a larger force than beach front buildings due to a funneling effect. These types of interactions are very poorly understood, and require additional research.

Hydrodynamic Modeling of Tsunami Evolution

Numerical simulations of tsunami propagation have made great progress in the last 30 years.



Tsunami Inundation, Modeling of, Fig. 3 Example of foundation scour. This image was taken by the International Tsunami Survey Team to Sri Lanka in January 2005

Several tsunami computational models are currently used in the National Tsunami Hazard Mitigation Program, sponsored by the National Oceanic and Atmospheric Administration, to produce tsunami inundation and evacuation maps for the states of Alaska, California, Hawaii, Oregon, and Washington. The computational models include MOST (Method Of Splitting Tsunami), developed originally by researchers at the University of Southern California (Titov and Synolakis 1998); COMCOT (Cornell Multi-grid Coupled Tsunami Model), developed at Cornell University (Liu et al. 1994); and TUNAMI-N2, developed at Tohoku University in Japan (Imamura 1995). All three models solve the same depth-integrated and 2D horizontal (2DH) nonlinear shallow-water (NSW) equations with different finite-difference algorithms. There are a number of other tsunami models as well, including the finite element model ADCIRC (ADvanced CIRCulation Model For Oceanic, Coastal And Estuarine Waters; e.g., Priest et al. 1997). For a given source region



Tsunami Inundation, Modeling of, Fig. 4 Example of damage to the backside of a coastal residence. These images were taken by the International Tsunami Survey Team to Sri Lanka in January 2005. The top photograph shows the front side of the structure, facing the ocean; there is damage but the main structure is intact. The lower photo shows the backside of the same building, showing the walls blown out, away from the center of the structure

condition, existing models can simulate propagation of a tsunami over a long distance with sufficient accuracy, provided that accurate bathymetry data exist.

The shallow-water equation models commonly lack the capability of simulating dispersive waves, which, however, could well be the dominating features in landslide-generated tsunamis and for the fission processes described previously. Several high-order depth-integrated wave hydrodynamics models (Boussinesq models) are now available for simulating nonlinear and weakly dispersive waves, such as COULWAVE (Cornell University Long and Intermediate Wave Modeling Package; Lynett and Liu 2002) and FUNWAVE (Kennedy et al. 2000). The major difference between the two is their treatment of moving shoreline boundaries. Lynett et al. (2003) applied COULWAVE to the 1998 PNG tsunami

with the landslide source; the results agreed with field survey data well. Recently, several finite element models have also been developed based on Boussinesq-type equations (e.g., Woo and Liu 2004). Boussinesq models require higher spatial and temporal resolutions, and therefore are more computationally intensive. Moreover, most of model validation work was performed for open-ocean or open-coast problems. In other words, the models have not been carefully tested for wave propagation and oscillations in semi-enclosed regions – such as a harbor or bay – especially under resonant conditions.

Being depth-integrated and horizontally 2D, NSW and Boussinesq models lack the capability of simulating the details of many coastal effects, such as wave overturning and the interaction between tsunamis and coastal structures, which could be either stationary or movable. At present, stationary coastal structures are parametrized as bottom roughness and contribute to frictional effects in these 2DH models. Although by adjusting the roughness and friction parameter satisfactory results can be achieved for maximum runup and delineation of the inundation zone (e.g., Liu et al. 1994), these models cannot provide adequate information for wave forces acting on coastal structures.

As a tsunami propagates into the nearshore region, the wave front undergoes a nonlinear transformation while it steepens through shoaling. It is in this nearshore region that dissipative effects can be important. Bottom friction can play a major role in the maximum runup and area of inundation (e.g. Tomita and Honda 2007). In depth-integrated models, bottom friction is typically approximated through a quadratic (drag) friction term, where the friction factor is calculated often through a Manning's coefficient or a Darcy–Wiesbach type friction factor (e.g. Kennedy et al. 2000; Liu et al. 1994). The validity of these steady-flow based coefficients has yet to be rigorously validated for use with tsunamis. If the tsunami is large enough, it can break at some offshore depth and approach land as a bore – the white wall of water commonly referenced by survivors of the Indian Ocean tsunami. Wave breaking in traditional NSW tsunami models has not

been handled in a satisfactory manner. Numerical dissipation is commonly used to mimic breaking (e.g. Liu et al. 1995), and thus results become grid dependant. In Boussinesq models, this breaking is still handled in an approximate manner due to the fact that the depth-integrated derivation does not allow for an overturning wave; however these breaking schemes have been validated for a wide range of nearshore conditions (e.g. Lynett 2006).

Being depth-integrated, NSW and Boussinesq models lack the capability of simulating the vertical details of many coastal effects, such as strong wave breaking/overturning and the interaction between tsunamis and irregularly shaped coastal structures. To address this deficiency, several 2D and 3D computational models based on Navier–Stokes equations have been developed, with varying degrees of success. An example is COBRAS (Cornell Breaking waves and structures model Lin and Liu 1998a, b; Lin et al. 1999), which is capable of describing the interactions between breaking waves and structures that are either surface piercing or submerged (Chang et al. 2001; Hsu et al. 2002). COBRAS adopted the Volume of Fluid (VOF) method to track free surface movement along with a Large Eddy Simulation (LES) turbulence closure model; several other computational models using different free surface tracking methods are also in use, such as the micro surface cell technique developed by Johnson et al. (1994). This 3D Navier–Stokes equation model has been tested by two tsunami related experiments. The first is 3D landslide experiments (Liu et al. 2005a, b), while the second involves measurements of solitary wave forces on vertical cylinders. Both experiments were conducted in the NEES tsunami basin at Oregon State. An example of a LES numerical solution of a solitary wave impinging on a circular is shown in Fig. 5.

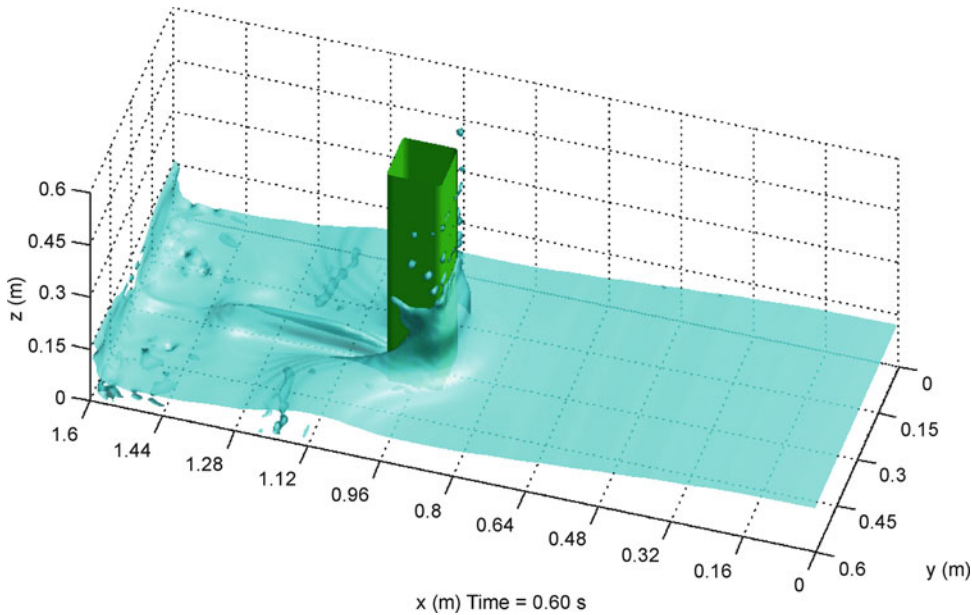
Due to their high computational costs, full 3D models would best be used in conjunction with a depth-integrated 2DH model (i. e., NSW or Boussinesq). While the 2DH model provides incident far-field tsunami information, the 3D model computes local wave-structure interactions. The results from 3D models could also provide a better parametrization of small-scale features (3D), which could then be embedded in a large-scale

2DH model. One-way coupling (e.g. using a NSW-generated time series to drive a 3D model, but not permitting feedback from the 3D model back into the NSW) is fairly straightforward to construct (e.g. Guignard et al. 1999). Two-way coupling, however, is difficult and requires consistent matching of physics and numerical schemes across model interfaces. Previous work in this area of two-way coupling of hydrodynamic models is limited. Fujima et al. (2002) two-way coupled a NLSW model with a fully 3D model. While the results appear promising, the approach used by Fujima et al. requires ad-hoc and unphysical boundary conditions at the model matching locations, in the form of spatial gradients forced to zero, to ensure numerical stability. Even with these ad-hoc treatments, their hybrid model compares very well with the completely-3D-domain simulation, requiring roughly 1/5 of the total 3D CPU time to achieve similar levels of accuracy. Sittanggang et al. (2006) presented work on two-way coupling of a Boussinesq model and 2D Navier–Stokes model. These results indicate that there is large potential for hybrid modeling, in terms of more rapid simulation as well as the ability to approach a new class of problems.

Moving Shoreline Algorithms

In order to simulate the flooding of dry land by a tsunami, a numerical model must be capable of allowing the shoreline to move in time. Here, the shoreline is defined as the spatial location where the solid bottom transitions from submerged to dry, and is a function of the two horizontal spatial coordinates and time. Numerical models generally require some type of special consideration and treatment to accurately include these moving boundaries; the logic and implementation behind this treatment is called a moving shoreline, or *runup*, algorithm.

For typical tsunami propagation models, it is possible to divide *runup* algorithms into two main approaches: those on a fixed grid and those on a Lagrangian or transformed domain. Both approaches have their advantages and



Tsunami Inundation, Modeling of, Fig. 5 An simulation snapshot taken from a 3D Navier–Stokes solver with an LES turbulence closure. This setup is looking at a bore

impacting a column; the wake and vertical splash are clearly visible. (Image provided by Philip L.-F. Liu, Cornell University)

disadvantages; currently fixed grid methods are found more commonly in operational-level models (e.g. Titov and Synolakis 1998), likely due in large part to their conceptual simplicity. A review of these two classes of models will be given in this section, followed by a review of the standard analytical, experimental, and field benchmarks used to validate the runup models. For additional information, the reader is directed to the comprehensive review given in Pedersen (2006).

With a fixed grid method, the spatial locations of the numerical grid points or control volumes are determined at the start of a simulation, and do not change shape or location throughout the simulation duration. These methods can be classified into extrapolation, stairstep, auxiliary shoreline point, and permeable beach techniques. The extrapolation method has its roots in Sielecki and Wurtele (1970), with extensions by Hibberd and Peregrine (1979), Kowalik and Murty (1993), and Lynett et al. (2002). The basic idea behind this method is that the shoreline location can be extrapolated using the nearest wet points, such that its position is not required to be locked onto

a fixed grid point; it can move freely to any location. Theoretically, the extrapolation can be of any order; however, from stability constraints a linear extrapolation is generally found. Hidden in the extrapolation, the method is roughly equivalent to the use of low-order, diffusive directional differences taken from the last wet point into the fluid domain (Lynett et al. 2002). Additionally, there are no explicit conservation constraints or physical boundary conditions prescribed at the shoreline, indicating that large local errors may result if the flow in the extrapolated region cannot be approximately as linear in slope. The extrapolation approach can be found in both NLSW and Boussinesq models with finite difference, finite volume, and finite element solution schemes, and has shown to be accurate for a wide range of non-breaking, breaking, two horizontal dimension, and irregular topography problems (e.g. Cheung et al. 2003; Cienfuegos et al. 2007; Korycansky and Lynett 2007; Lynett et al. 2003; Pedrozo-Acuna et al. 2006).

Stairstep moving shoreline methods, one of the more common approaches found in tsunami models (e.g. Liu et al. 1994), reconstruct the

naturally continuous beach profile into a series of constant elevation segments connected through vertical transitions. In essence, across a single cell width, the bottom elevation is taken as the average value. A cell transitions from a dry cell to a wet cell when the water elevation in a neighboring cell exceeds the bottom elevation, and transitions from wet to dry when the local total water depth falls below some small threshold value. These methods are particularly useful in finite volume and C-grid (Arakawa and Lamb 1977) type approaches (e.g. LeVeque and George 2004; Liu et al. 1995), but can be difficult to implement in centered difference models, particularly high-order models or those sensitive to fluid discontinuities, where the “shock” of opening and closing entire cells can lead to numerical noise.

Auxiliary shoreline point methods require dynamic re-gridding very near the shoreline, such that the last wet point is always located immediately at the shoreline. Obviously, this method requires a numerical scheme that can readily accommodate non-uniform and changing node locations. There is some relation to the extrapolation methods discussed above; the moving shoreline point must be assigned some velocity, and it is extrapolated from the neighboring wet points. However, it is fundamentally different in that the shoreline point is explicitly included in the fluid domain. Thus, it would be expected that the governing conservation equations near the shoreline are more precisely satisfied here, although still dependent on the appropriateness of the extrapolation. One such method can be found in Titov and Synolakis (1995), and has been successfully applied in NSLW equation models.

Another fixed grid treatment of moving boundary problems is employing a slot or permeable-seabed technique (Tao 1983, 1984). Conceptually, this method creates porous slots, or conduits, through the dry beach, such that there is always some fluid in a “dry” beach cell, although it may exist below the dry beach surface. These porous, “dry” nodes use a modified form of the NLSW; it is noted here that although in concept this approach is modeling a porous beach, it is not attempting to simulate the groundwater flow under a real, sandy beach, for example. The

equations governing the “dry” domain contain a number of empirical parameters that are tuned to provide reasonable runup agreement with benchmark datasets. The advantage of this approach is that it allows the entire domain, including the fluid and “dry” nodes, to be determined via a somewhat consistent set of governing equations, without requiring a direct search routine to determine the shoreline location. The method has gained some popularity in wind wave models (e.g. Kennedy et al. 2000; Madsen et al. 1997) when a highly accurate estimate of the shoreline location is not the highest priority. However, the approach has been used with some success in tsunami studies (e.g. Ioualalen et al. 2007) despite the fact that the empirical coefficients that govern the model accuracy cannot be universally determined for a wide range of problems (Chen et al. 2000).

Alternative to fixed grid methods is the Lagrangian approach. Here, the fluid domain is discretized into particles, or columns of fluid in depth-integrated models, that are transported following the total fluid derivative. There are no fixed spatial grid locations; the columns move freely in space and time and thus these techniques require numerical flexibility, in terms of utilizing constantly changing space and time steps. The Lagrangian approach can be described as both the more physically consistent and mathematically elegant method of describing shoreline motion. The shoreline “particle” is included in the physical formulation just as any other point in the domain (i. e. no extrapolations are necessary), and thus the shoreline position accuracy will be compromised only by the overarching physical approximation (e.g. long wave approximation) and the numerical solution scheme (e.g. second-order time integration). The cost for this accuracy is a mathematical system that can be more difficult and tedious to solve numerically, typically requiring domain transformations, mappings, and/or re-griddings. Lagrangian methods have been used successfully in finite difference and finite element nonlinear shallow water (NLSW) and Boussinesq equation models (e.g., Birknes and Pedersen 2006; Gopalakrishnan and Tung 1983; Özkan-Haller and Kirby 1997; Pedersen and

Gjevik 1983; Petera and Nassehi 1996; Zelt 1991).

Future Directions

Towards a more robust simulation of tsunami inundation, there are two major issues which require additional fundamental investigations: dissipation mechanisms and interaction with infrastructure. Bottom friction, known to play an important role in inundation, needs to be re-examined starting from its basic formulation. Can a steady-flow based Mannings-type expression for bottom friction be used for tsunami? Does the unsteady nature of the tsunami flow make use of these existing formulations invalid? The answer to these questions may be different depending on what part of the wave is investigated (e.g. front). In addition, the hydrodynamic effect of common coastal vegetation, such as mangroves, needs to be quantified. There is current discussion of the use of such natural roughness as a tsunami defense (e.g. Danielsen et al. 2005); confidence cannot be put in such measures until it is understood how they behave. In addition to bottom friction, which exists at all locations and times under an inundating tsunami, wave breaking can increase the total energy dissipation. While breaking is generally confined to the leading front of a tsunami, the characteristics of this front are important for hydrodynamic loadings on beachfront structures, and may be significant to the net sediment and debris transport of a tsunami. Three-dimensional tsunami breaking is poorly understood and has received little attention.

Wave loadings and interactions with infrastructure are not well understood. To tackle this problem, tsunami hydrodynamic models need to be coupled with structural and geotechnical models. Ideally, these models should all be two-way coupled, such that the displacement of a structure, be it a single collapsed wall, will change the flow pattern, and scour underneath the foundation will change the structure stability. Additionally, impacts of flow-transported debris (e.g. cars) should be included in this framework. If such a modeling capacity existed, engineering

design of coastal structures could be undertaken in a very efficient manner.

Bibliography

- Arakawa A, Lamb VR (1977) Computational design of the basic dynamical processes of the UCLA general circulation model. In: Chang J (ed) *Methods in computational physics*. Academic, New York, pp 174–267
- Birknes J, Pedersen G (2006) A particle finite element method applied to long wave run-up. *Int J Numer Methods Fluids* 52(3):237–261
- Borrero JC, Bourgeois J, Harkins G, Synolakis CE (1997) How small-scale bathymetry affected coastal inundation in the 1992 Nicaraguan tsunami. Fall AGU Meeting, San Francisco
- Briggs MJ, Synolakis CE, Harkins GS, Green D (1994) Laboratory experiments of tsunami runup on a circular island. *Pure Appl Geophys* 144(3/4):569–593
- Carrier GF (1966) Gravity waves on water of variable depth. *J Fluid Mech* 24:641–659
- Chang K-A, Hsu T-J, Liu PL-F (2001) Vortex generation and evolution in water waves propagating over a submerged rectangular obstacle. Part I solitary waves. *Coast Eng* 44:13–36
- Chen Q, Kirby JT, Dalrymple RA, Kennedy AB, Chawla A (2000) Boussinesq modeling of wave transformation, breaking, and runup: Part I 2D. *J Waterw Port Coast Ocean Eng* 126(1):57–62
- Cheung K, Phadke A, Wei Y, Rojas R, Douyere Y, Martino C, Houston S, Liu PL-F, Lynett P, Dodd N, Liao S, Nakazaki E (2003) Modeling of storm-induced coastal flooding for emergency management. *Ocean Eng* 30:1353–1386
- Cienfuegos R, Barthelemy E, Bonneton P (2007) A fourth-order compact finite volume scheme for fully nonlinear and weakly dispersive Boussinesq-type equations. Part II: boundary conditions and model validation. *Int J Numer Methods Fluids* 53(9):1423–1455
- Cross RH (1967) Tsunami surge forces, ASCE JI Waterways & Harbors Division WW4:201–231
- Danielsen F, Sørensen MK, Olwig MF, Selvam V et al (2005) The Asian tsunami: a protective role for coastal vegetation. *Science* 310:643
- Fernando HJS, McCulley JL, Mendis SG, Perera K (2005) Coral poaching worsens tsunami destruction in Sri Lanka. *Eos Trans AGU* 86(33):301
- Fujima K, Masamura K, Goto C (2002) Development of the 2d/3d hybrid model for tsunami numerical simulation. *Coast Eng J* 44(4):373–397
- Geist E, Lynett P, Chaytor J (2008) Hydrodynamic modeling of tsunamis from the Currituck landslide, Marine Geology (in press)
- González FI, Satake K, Boss EF, Mofjeld HO (1995) Edge wave and non-trapped modes of the 25 April 1992 Cape Mendocino tsunami. *Pure Appl Geophys* 144(3–4):409–426

- Gopalakrishnan TC, Tung CC (1983) Numerical analysis of a moving boundary problem in coastal hydrodynamics. *Int J Numer Methods Fluids* 3:179–200
- Goring DG, Raichlen F (1992) Propagation of long waves onto shelf. *J Waterw Port Coast Ocean Eng* 118(1): 43–61
- Guignard S, Grilli ST, Marcer R, Rey V (1999) Computation of shoaling and breaking waves in nearshore areas by the coupling of BEM and VOF methods. In: Proceedings of the 9th offshore and polar engineering conference, vol III, ISOPE99, Brest, pp 304–309
- Hammack J, Segur H (1978) Modelling criteria for long water waves. *J Fluid Mech* 84(2):359–373
- Hibberd S, Peregrine DH (1979) Surf and run-up on a beach. *J Fluid Mech* 95:323–345
- Horrrillo J, Kowalik Z, Shigihara Y (2006) Wave dispersion study in the Indian Ocean–Tsunami of December 26, 2004. *Mar Geod* 29(3):149–166(18)
- Hsu T-J, Sakakiyama T, Liu PL-F (2002) A numerical model for waves and turbulence flow in front of a composite breakwater. *Coast Eng* 46:25–50
- Imamura F (1995) Review of tsunami simulation with a finite difference method, long-wave runup models. World Scientific, Singapore, pp 25–42, http://en.wikipedia.org/wiki/2004_Indian_Ocean_earthquake
- Ioualalen M, Asavanant JA, Kaewbanjak N, Grilli ST, Kirby JT, Watts P (2007) Modeling of the 26th December 2004 Indian Ocean tsunami: case study of impact in Thailand. *J Geophys Res* 112:C07024. <https://doi.org/10.1029/2006JC003850>
- Johnson DB, Raad PE, Chen S (1994) Simulation of impacts of fluid free surfaces with solid boundaries. *Int J Numer Methods Fluids* 19:153–176
- Johnson RS (1972) Some numerical solutions of a variable-coefficient Korteweg–de Vries equation (with application to solitary wave development on a shelf). *J Fluid Mech* 54:81
- Kennedy AB, Chen Q, Kirby JT, Dalrymple RA (2000) Boussinesq modeling of wave transformation, breaking, and runup. 1:1D. *J Waterw Port Coast Ocean Eng* 126:39–47
- Korycansky DG, Lynett P (2007) Runup from impact tsunami. *Geophys J Int* 170:1076–1088
- Kowalik Z, Murty TS (1993) Numerical simulation of two-dimensional tsunami runup. *Mar Geod* 16:87–100
- Kulikov E (2005) Dispersion of the Sumatra tsunami waves in the Indian Ocean detected by satellite altimetry. Report from P.P. Shirshov Institute of Oceanology, Russian Academy of Sciences, Moscow
- Lay T, Kanamori H, Ammon C, Nettles M, Ward S, Aster R, Beck S, Bilek S, Brudzinski M, Butler R, DeShon H, Ekström G, Satake K, Sipkin S (2005) The great Sumatra–Andaman earthquake of December 26, 2004. *Science* 308:1127–1133. <https://doi.org/10.1126/science.1112250>
- LeVeque RJ, George DL (2004) High-resolution finite volume methods for the shallow water equations with bathymetry and dry states. In: Liu PL, Yeh H, Synolakis C (eds) Advanced numerical models for simulating tsunami waves and runup. Advances in coastal and ocean engineering, vol 10. World Scientific, Singapore
- Liu PL-F, Cheng Y (2001) A numerical study of the evolution of a solitary wave over a shelf. *Phys Fluids* 13(6): 1660–1166
- Liu PL-F, Synolakis CE, Yeh H (1991) Report on the international workshop on long-wave run-up. *J Fluid Mech* 229:675–688
- Liu PL-F, Cho Y-S, Yoon SB, Seo SN (1994) Numerical simulations of the 1960 Chilean tsunami propagation and inundation at Hilo, Hawaii. In: El-Sabh MI (ed) Recent development in tsunami research. Kluwer, Dordrecht, pp 99–115
- Liu PL-F, Cho Y-S, Briggs MJ, Kanoglu U, Synolakis CE (1995) Runup of solitary waves on a circular island. *J Fluid Mech* 320:259–285
- Liu PL-F, Lynett P, Fernando H, Jaffe BE, Fritz H, Higman B, Morton R, Goff J, Synolakis C (2005a) Observations by the International Tsunami Survey Team in Sri Lanka. *Science* 308:1595
- Liu PL-F, Wu T-R, Raichlen F, Synolakis CE, Borrero JC (2005b) Runup and rundown generated by three-dimensional sliding masses. *J Fluid Mech* 536: 107–144
- Losada MA, Vidal V, Medina R (1989) Experimental study of the evolution of a solitary wave at an abrupt junction. *J Geophys Res* 94:14557
- Lynett P (2006) Nearshore modeling using high-order Boussinesq equations. *J Waterw Port Coast Ocean Eng* 132(5):348–357
- Lynett P (2007) The effect of a shallow water obstruction on long wave runup and overland flow velocity. *J Waterw Port Coast Ocean Eng* 133(6):455–462
- Lynett P, Liu PL-F (2002) A numerical study of submarine landslide generated waves and runup. *Proc R Soc Lond A* 458:2885–2910
- Lynett P, Wu T-R, Liu PL-F (2002) Modeling wave runup with depth-integrated equations. *Coast Eng* 46(2): 89–107
- Lynett P, Borrero J, Liu PL-F, Synolakis CE (2003) Field survey and numerical simulations: a review of the 1998 Papua New Guinea tsunami. *Pure Appl Geophys* 160: 2119–2146
- Madsen OS, Mei CC (1969) The transformation of a solitary wave over an uneven bottom. *J Fluid Mech* 39:781
- Madsen PA, Sorensen OR, Schaffer HA (1997) Surf zone dynamics simulated by a Boussinesq-type model: part I. Model description and cross-shore motion of regular waves. *Coast Eng* 32:255–287
- Matsuyama M, Ikeno M, Sakakiyama T, Takeda T (2007) A study on tsunami wave fission in an undistorted experiment. *Pure Appl Geophys* 164:617–631
- Özkan-Haller HT, Kirby JTA (1997) Fourier-Chebyshev collocation method for the shallow water equations including shoreline run-up. *Appl Ocean Res* 19: 21–34
- Pedersen G (2006) On long wave runup models. In: Proceedings of the 3rd international workshop on long-wave runup models, Catalina Island

- Pedersen G, Gjevik B (1983) Runup of solitary waves. *J Fluid Mech* 142:283–299
- Pedrozo-Acuna A, Simmonds DJ, Otta AK, Chadwick AJ (2006) On the cross-shore profile change of gravel beaches. *Coast Eng* 53(4):335–347
- Petera J, Nassehi V (1996) A new two-dimensional finite element model for the shallow water equations using a Lagrangian framework constructed along fluid particle trajectories. *Int J Numer Methods Eng* 39:4159–4182
- Priest GR et al (1997) Cascadia subduction zone tsunamis: hazard mapping at Yaquina Bay, Oregon. Final Technical Report to the National Earthquake Hazard Reduction Program DOGAMI Open File Report 0-97-34, p 143
- Ramsden JD (1996) Forces on a vertical wall due to long waves, bores, and dry-bed surges. *J Waterw Port Coast Ocean Eng* 122(3):134–141
- Seabra-Santos FJ, Renouard DP, Temperville AM (1987) Numerical and experimental study of the transformation of a solitary wave over a shelf or isolated obstacles. *J Fluid Mech* 176:117
- Shen M, Meyer R (1963) Climb of a bore on a beach Part 3. Runup. *J Fluid Mech* 16(1):113–125
- Sielecki A, Wurtele MG (1970) The numerical integration of the nonlinear shallow-water equations with sloping boundaries. *J Comput Phys* 6:219–236
- Sittanggang K, Lynett P, Liu P (2006) Development of a Boussinesq-RANSVOF hybrid wave model. In: Proceedings of 30th ICCE, San Diego, pp 24–25
- Synolakis CE (1987) The runup of solitary waves. *J Fluid Mech* 185:523–545
- Synolakis CE, Imamura F, Tsuji Y, Matsutomi S, Tinti B, Cook B, Ushman M (1995) Damage, conditions of East Java tsunami of 1994 analyzed. *EOS Trans Am Geophys Union* 76(26):257, 261–262
- Tadepalli S, Synolakis CE (1994) The runup of N-Waves on sloping beaches. *Proc R Soc Lond A* 445:99–112
- Tadepalli S, Synolakis CE (1996) Model for the leading waves of tsunamis. *Phys Rev Lett* 77:2141–2144
- Tao J (1983) Computation of wave runup and wave breaking. Internal report. Danish Hydraulics Institute, Denmark
- Tao J (1984) Numerical modeling of wave runup and breaking on the beach. *Acta Oceanol Sin* 6(5): 692–700. (in chinese)
- Titov VV, Synolakis CE (1995) Modeling of breaking and nonbreaking long wave evolution and runup using VTCS-2. *J Waterw Port Coast Ocean Eng* 121(6): 308–316
- Titov VV, Synolakis CE (1998) Numerical modeling of tidal wave runup. *J Waterway Port Coast Ocean Eng* 124(4):157–171
- Tomita T, Honda K (2007) Tsunami estimation including effect of coastal structures and buildings by 3D model. Coastal Structures '07, Venice
- Ward SN, Day S (2001) Cumbre Vieja Volcano – potential collapse and tsunami at La Palma, Canary Islands. *Geophys Res Lett* 28(17):3397–3400
- Weiss R, Wunnenmann K, Bahlburg H (2006) Numerical modelling of generation, propagation, and run-up of tsunamis caused by ocean impacts: model strategy and technical solutions. *Geophys J Int* 67:77–88
- Woo S-B, Liu PL-F (2004) A finite element model for modified Boussinesq equations. Part I: model development. *J Waterway Port Coast Ocean Eng* 130(1):1–16
- Yeh H (2006) Maximum fluid forces in the tsunami runup zone. *J Waterway Port Coast Ocean Eng* 132(6):496–500
- Yeh H, Ghazali A, Marton I (1989) Experimental study of bore runup. *J Fluid Mech* 206:563–578
- Yeh H, Liu P, Briggs M, Synolakis C (1994) Propagation and amplification of tsunamis at coastal boundaries. *Nature* 372:353–355
- Zelt JA (1991) The run-up of nonbreaking and breaking solitary waves. *Coast Eng* 15(3):205–246

Tsunami Sedimentology

Pedro J. M. Costa¹ and S. Dawson²

¹Instituto D. Luiz and Departamento de Geologia da Universidade de Lisboa, Faculdade de Ciências da, Universidade de Lisboa, Lisbon, Portugal

²Geography, School of Social Sciences, University of Dundee, Dundee, Scotland, UK

Article Outline

Glossary
Definition
Introduction
Description
Future Directions
Bibliography

Glossary

Closure depth Is a key parameter in coastal processes and marks the limit where waves interact with sea bottom sediments.

Cross-lamination An arrangement of strata that are locally inclined at some angle to the overall planar orientation of the stratification.

Heavy minerals Minerals with a density above 2.9 g/cm³.

Laminae Thin layer of sediment or sedimentary rock.

Loading structures Deformation sediment structure formed during soft-sediment deformation due to overlying weight.

Microtextures Microscopic imprints in the surface of sediment grains. Typically analyzed after observation under the scanning electron microscope that allows significant magnifications of the grains.

Mud drapes Sedimentation of fine sediments on top of coarser layers occurring only under low-flow conditions.

Normal grading Fining-upward sequence with coarser sediments at the base progressively becoming finer to the top.

Parallel lamination or horizontal lamination Unit or layer typically with a horizontal base and parallel overlaying *laminae*.

Rip-up clasts (Typically fine) material eroded from the underlying layers and incorporated in the (tsunamigenic or storm) deposits.

Sediment structures Macroscopic three-dimensional features of sedimentary rocks or sediments recording processes occurring during deposition or between deposition and lithification. Their recognition and application are relevant in the definition of depositional environments, fabric, history, or surface processes. Primary sedimentary structures occur in clastic sediments and produced by the same processes (waves, currents, etc.) that caused deposition and includes plane bedding and cross-bedding. Secondary sedimentary structures are caused by postdepositional processes, including biogenic, chemical, and mechanical disruption of sediment.

Sedimentology Field of earth sciences studying sediments and/or sedimentary rocks and the processes responsible for their formation.

Tsunami Gravity wave system following a large-scale disturbance of the sea surface. This disturbance is caused by a vertical displacement of the water column as a result of an earthquake, landslide, volcanic event, or meteor impact. Tsunami waves are characterized by its long wave length and high velocity (in deeper ocean).

Tsunami deposit Sedimentary evidence detected in coastal stratigraphy resulting from tsunami inundation and its associated sediment transport and deposition.

Definition

Tsunami sedimentology is a field of Science devoted to the study of the sedimentary features of tsunami deposits. Tsunami inundation (and its associated backwash) promotes changes in the sedimentary balance of a coastal area, typically promoting erosion in the coastal fringe and deposition farther inland. Tsunami sedimentology applies a wide range of sedimentological techniques to identify and differentiate these units in the coastal stratigraphy. These techniques provide data reflecting the characteristics of tsunami waves, transport mechanisms, preservation potential, and sedimentary sources. The identification of tsunami deposits enables estimation of inundation distance and run-up, although patterns of erosion and deposition by both landward- and seaward-directed flows introduce uncertainties in those reconstructions. Examination of palaeotsunami deposits leads to inferences on the recurrence intervals of extreme events in the coastal zone and improves hazard planning and mapping.

Introduction

Tsunami deposition is generally characterized by the entrainment and redeposition of shallow marine or coastal sediments in terrestrial and/or transitional (e.g., back-barrier, lagoonal, estuarine) sedimentary environments. The recognition of tsunami signatures is frequently restricted by poor preservation (or absence) of those deposits in the stratigraphic record. In many cases, the erosional characteristics of the event, the absence of lithological and textural contrasts between exotic and local sediments, anthropogenic activity, and natural postdepositional disturbances eliminate or make palaeotsunami deposits difficult to identify; as a result, it is often difficult to state the return intervals of such events (e.g., Szczucinski 2012). Thus, the accurate assessment of extreme marine inundation hazards by tsunami for any given coastal area requires precise identification and differentiation of tsunami sedimentary deposits, and this can only be achieved from the use of a range of sedimentological proxies

(see, e.g., Chagué-Goff et al. 2011). Despite the multiproxy approach, the catalogues derived from the geological record will probably be biased toward the largest events and incomplete in the number of events recorded.

The first studies to use the geological record in detecting prehistoric tsunamis were conducted by Atwater (1986) and Dawson et al. (1988). Since then many papers have been published discussing sedimentological features of tsunami deposits and associating those features to transport and depositional processes. In particular, the study of modern, well-preserved deposits carried out during post-tsunami surveys, following events across the globe in the last two decades, exclude uncertainties on the generating event and minimize natural and anthropogenic disturbances, provided the opportunity to refine palaeotsunami diagnostic criteria. During the last two decades, several authors (e.g., Chagué-Goff et al. 2011; Dawson and Stewart 2007; Gelfenbaum and Jaffe 2003) have postulated criteria to distinguish (palaeo)tsunami deposits.

In contrast to modern tsunami, for which eyewitness accounts and field measurements of both erosional and depositional effects are utilized in modeling studies, (palaeo)tsunami recognition depends on the identification of ancient tsunami deposits. Recognition of these deposits is the primary method for reconstructing tsunami minimum inundation distance and run-up, although patterns of erosion and deposition by both landward- and seaward-directed flows are complex, these patterns being further complicated by the existence of more than one wave associated with the same tsunami (Bondevik et al. 1997; Paris et al. 2007), thus introducing uncertainties in those reconstructions.

The nature of tsunami deposits varies greatly with coastal and nearshore morphology, the height of tsunami waves at the coast and run-up, and with the nature and amount of existing sediment in any coastal setting when affected by such an event. Consequently, the possible variations in sedimentary processes and associated deposits during these complex events remain poorly understood, but in general a tsunami deposit will only be produced if there is a suitable supply of sediment and accommodation space in the

coastal zone. More recently, the subsequent backwash, or return flow, has been regarded as a process of significant geomorphic and sedimentologic consequences (Dawson 1994; Dawson 1999), though the spatial extension of the correspondent signature is usually more restricted due to channeling effects. The geomorphological consequences of tsunamis and storms in coastal dunes or barrier islands have also been addressed as the differentiation between both types of event is often problematic in palaeo-coastal reconstruction (e.g., Andrade 1992; Goff et al. 2010b).

The specific sediment transport processes and physics of tsunami and storm wave inundation tend to leave their sediment imprint in a wide range of environments (e.g., alluvial plains, estuaries, coastal lagoons, embayments, nearshore and offshore areas) although storms usually present a smaller amount of inland penetration. However, many of these environments display low preservation potential for event deposits, and the recognition of tsunami and storm deposits is often constrained by the poor preservation of those deposits (or absence) in the stratigraphic record.

Description

Tsunami Deposits

The identification and differentiation of the more finer units deposited by tsunami or storms in coastal stratigraphy require a multiproxy approach that mainly focuses on the allochthonous sediment and/or palaeontological content in order to establish a marine or coastal provenance. Tsunami-transported sediment is typically deposited during run-up, even though deposition also occurs during backwash and in the period of time between run-up and backwash, usually correspondent to a slack where currents fall to minimum intensity and the directional pattern is ill defined (Fig. 1). Furthermore, in some situations the erosional capacity of run-up or the backwash constrains the sedimentary recognition of events by removing sediments potentially deposited by earlier tsunami waves.

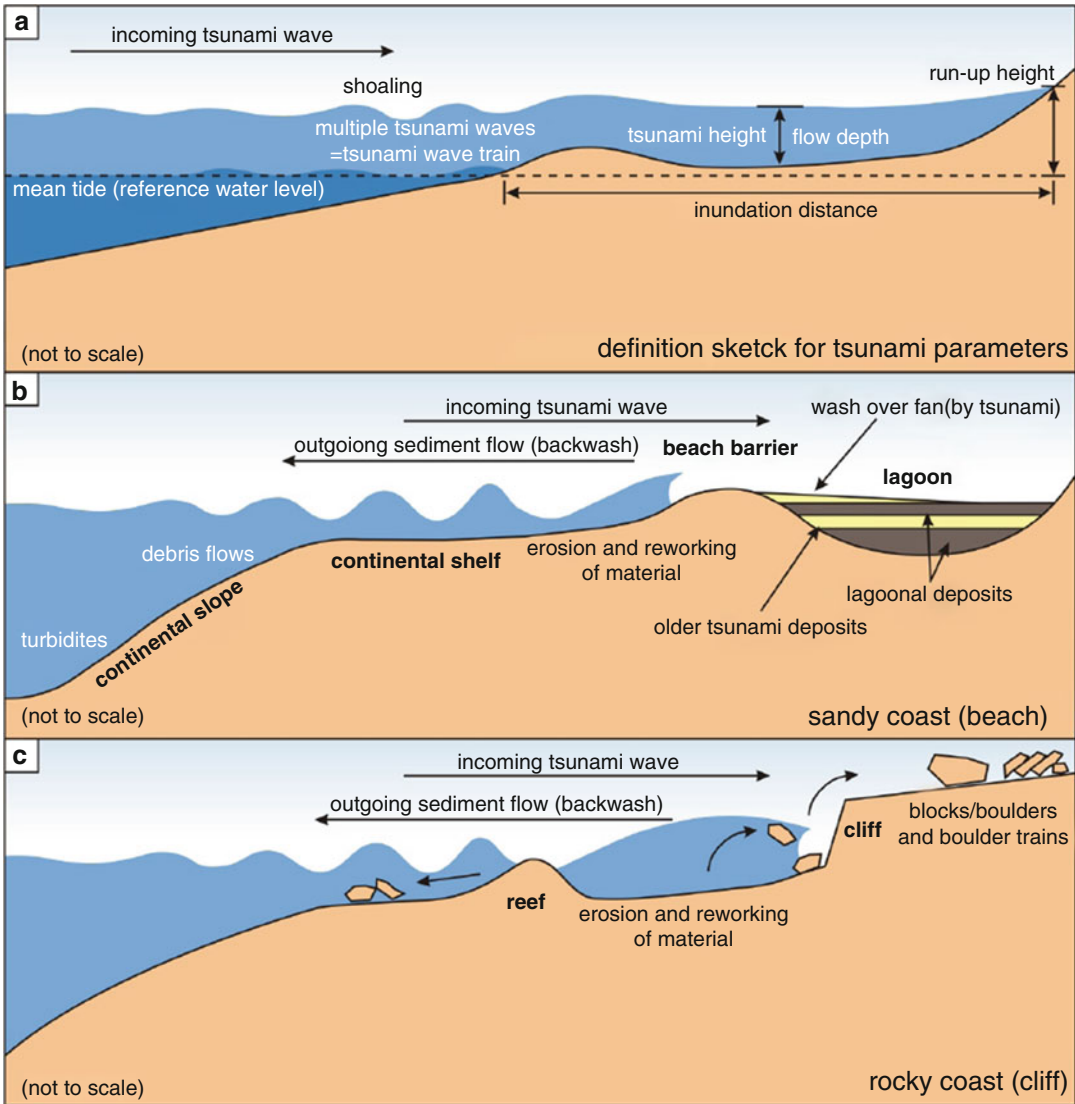
There are two main types of onshore sedimentary evidences associated with tsunami and

storms: deposition of large boulders and deposition of finer (typically sand-sized) sediments in coastal areas. The universally accepted scale of grain sizes of siliciclastic sediment is the Udden–Wentworth scale that is composed of four basic groups: clay ($<4\ \mu\text{m}$), silt ($4\text{--}63\ \mu\text{m}$), sand ($63\text{--}2\ \text{mm}$), and gravel ($>2\ \text{mm}$). The larger particles (i.e., pebbles, cobbles, and boulders) were all included in the gravel group.

Boulder and finer deposits can provide clues to the identification of tsunami events. In the case of larger clasts, the differentiation is firstly based on the identification of boulders that have been transported inland or upward from or within the coastal zone and against gravity. In some cases, these boulders appear simply overturned a few m inland from their original source area. The recognition of boulder deposits associated with both tsunami and storms has been intensely debated in the literature (Goff et al. 2010b; Goto et al. 2010a; Goto et al. 2010b; Nandasena et al. 2011; Nott 1997; Scheffers and Kelletat 2005; Scheffers and Scheffers 2007; Scheffers et al. 2008). From the many examples in the literature, a few deserve special notice because of their specific lithological, geological, geomorphological, or oceanographic significance. The majority of tsunami sediments identified within coastal stratigraphy are coarser sand-sized layers within low-energy finer materials accumulated in depositional basins and exhibiting a distinctive sedimentological signature within the stratigraphic column (Fig. 2 and, e.g., Bondevik et al. 1997; Chagué-Goff et al. 2002; Dawson et al. 1988; Moore et al. 2007; Nanayama et al. 2007; Nanayama et al. 2000; Paris et al. 2009). This type of depositional arrangement is indicative of extreme marine inundations because they provide a stratigraphic context and facilitate accurate identification and dating of individual events. Fine tsunami or storm deposits are typically sand sized with a clearly defined clay/silt fraction (Srinivasalu et al. 2007).

Sedimentological Features of Onshore Tsunami Deposits

During the last two decades, several authors (e.g., Chagué-Goff et al. 2011; Dawson and Stewart 2007; Gelfenbaum and Jaffe 2003) have



Tsunami Sedimentology, Fig. 1 Overview of tsunami characteristics while inflow and backflow: (a) general definition of tsunami parameters and depositional

characteristics of tsunami deposits on (b) sandy coasts with lagoons or marshlands as well as (c) on rocky coasts or cliffs (From Koster 2014)

postulated criteria to distinguish (palaeo)tsunami deposits. These are described below and summarized in Table 1. Dawson and Stewart (2007) discussed the processes of tsunami deposition, identifying the three main aspects that make the depositional process unique, tsunami source, propagation, and inundation. The establishment of source material has been widely used (e.g., Costa et al. 2009; Dahanayake and Kulasena 2008; Dawson et al. 1996a; Gelfenbaum and

Jaffe 2003; Hindson et al. 1996; Minoura et al. 1996; Morton et al. 2007; Narayana et al. 2007; Paris et al. 2009; Szczucinski et al. 2006) because it allows one to reconstruct the origin and pathway of former tsunami waves. However, it has been commonly reported that tsunami waves transport essentially sediment that is available in the coastal fringe landward of the boundary defined by the seasonal depth of closure of the beach (and coastal) profile (e.g., Costa et al.



Tsunami Sedimentology, Fig. 2 Photographs exhibiting the AD 1755 sandy tsunami deposit in Portugal from several locations. Please note the sticking contrast with the permanent regime sedimentation pattern (alluvial or lagoonal muds) and the high energy associated with the deposition of sand (it is easy to observe the erosional basal

contact and the sharp top contact of the tsunamigenic unit). Image (a) Boca do Rio (vertical scale approx. 1 m). (b) Martinhal (vertical scale approx. 0.5 m). (c and e) Salgados (vertical scale approx. 0.5 m in the trench wall and 0.3 m in the box-core). (d) Alcantarilha (vertical scale approx. 0.70 m)

2009; Costa et al. 2012b; Dawson 1994; Goff et al. 2010a; Kortekaas and Dawson 2007; Paris et al. 2009). In contrast with this, micropalaeontological evidences have indicated either relevant changes in the population of nannoplankton, foraminifera, diatoms, and ostracods or that marine species from offshore/nearshore have been transported inland and deposited by tsunami (e.g., Clague et al. 2000; Dawson et al. 1996b; Hindson et al. 1996; Mamo et al. 2009; Sawai

et al. 2009; Shennan et al. 1996). Although a site-specific component might be a central feature of any tsunami deposits, some generalizations are possible interrelated with sedimentary structures, sediment source, and palaeontological, geochemical, and geomorphological signatures (Table 1).

Typically, tsunamis can leave sedimentary imprints on shores far from the event source and usually less than a kilometer from the coastline (e.g., Shi and Smith 2003). Tsunami deposits are

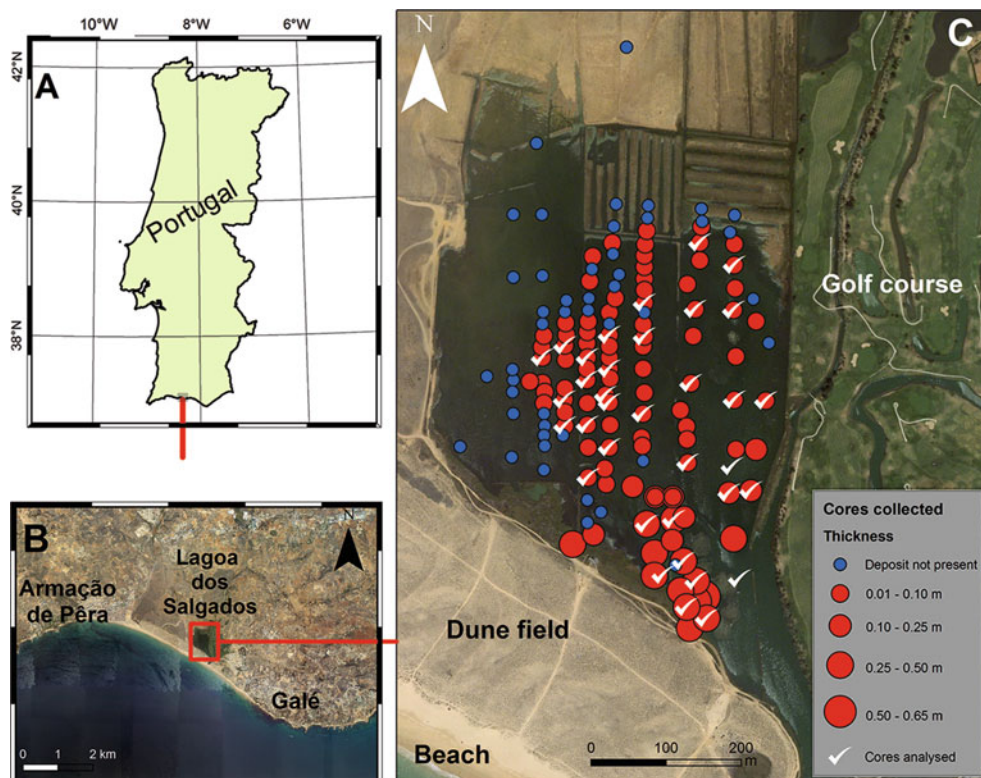
Tsunami Sedimentology, Table 1 Sedimentological criteria commonly used in identification of tsunami deposits

Criteria	Features
Sedimentary structures/features	Erosional/abrupt/sharp/unconformity basal contact Massive/chaotic Normal grading (in places repeated) Unit with several <i>laminae</i> Cross-stratification Parallel lamination or cross-lamination Soft-sediment deformation Loading structures Mud drapes Rip-up clasts Broken shells
Sediment source	Typically reflects the material available in the coastal fringe (i.e., beach, aeolian, inner shelf landward of closure depth) Grain size range from mud to boulders Multimodal grain-size distribution indicating multiple sources Increase of heavy mineral concentration in the base of the deposit Increase of platy minerals (i.e., micas) at the top of the deposit SEM microtextural imprints suggest increased presence of percussion/mechanic marks
Palaeontological features	Marked changes in diatoms, foraminifera, ostracods, nannoplankton, plant macrofossils
Geochemical signatures	Increase Cl, Na, Mg, Ca, K, SiO ₂ , CaO, Cr, MgO, I, Fe, S Increases in the ratios of SiO ₂ /Al ₂ O ₃ and CaO/Al ₂ O ₃ Increase in carbonate content (shell)

usually thicker in topographic lows (areas of spatial deceleration of flows) and thin over topographic highs (areas of spatial acceleration of flows). In fact tsunami sediments can also be eroded during phases of backwash and have also been linked to new phases of sedimentation during backwash.

Several authors have argued that tsunamis are frequently associated with the deposition of continuous or discontinuous sediment sheets across large areas of the coastal zone, provided that there is an adequate sediment supply (e.g., Dawson et al. 1996b; Hindson et al. 1996). The decrease of energy associated with the tsunami inundation is evidenced in stratigraphic and sedimentary architecture by the fining inland and thinning inland and ramping upward of tsunamigenic deposits (Fig. 3). This is probably the most common feature/criteria to recognize tsunami events in the stratigraphy of any given coastal area mainly due to the settlement of the particles through the water column, related to a decrease of the turbulence of the flow, generally forming

fining-upward depositional sequences. Grain size characteristics of the tsunami deposits reflect both the origin of the displaced sediment and hydrodynamic conditions of sedimentation (Sugawara et al. 2008), with normally graded sand layers related to the decrease of the hydrodynamic energy during sedimentation (e.g., Dawson et al. 1988; Minoura et al. 1996). Although not a frequent situation, each fining-upward sequence can be attributed to individual tsunami waves. Although local topography plays a decisive role (e.g., Hori et al. 2007), the thickness and mean grain size of tsunami deposits generally decrease landward (e.g., Gelfenbaum and Jaffe 2003; Goff et al. 2004; Minoura et al. 1996; Paris et al. 2009). Landward coarsening deposits have also been observed on rare occasions (e.g., Higman and Bourgeois 2008). In fact, the sediment texture of tsunami deposits is mostly related to material available for transport in the coastal zone, and the associated tsunami deposits can vary immensely from location to location. Differences in the tsunami records preserved tend to



Tsunami Sedimentology, Fig. 3 Image representing the decreasing thickness of a tsunamigenic deposit in a lowland in the Algarve coast (Portugal) (Adapted from Costa 2012)

reflect the unique character of each tsunami and may be attributed to source differences, coastal configuration, tide level, and sediment supply. For example, tsunami sediment source has been attributed to beaches and berms (e.g., Gelfenbaum and Jaffe 2003), aeolian grains (e.g., Costa et al. 2009), or to the inner shelf (e.g., Paris et al. 2007).

The use of heavy minerals to establish provenance of tsunamigenic deposits has also been investigated by several authors (e.g., Switzer et al. 2005; Bahlburg and Weiss 2006; Szczucinski et al. 2006; Babu et al. 2007; Morton et al. 2007; Narayana et al. 2007; Higman and Bourgeois 2008; Morton et al. 2008; Switzer and Jones 2008; Costa et al. 2015). For example, Bahlburg and Weiss (2006) observed the presence of concentrations of thin heavy mineral at the base of individual sand layers inferred to have been laid down by different

waves from the same event. Morton et al. (2008) observed that vertical textural trends showed an overall but nonsystematic upward fining and upward thinning of depositional units with an upward increase in heavy mineral laminations at some locations. However, most of these studies were limited to one study area and also by local differences in source material (Jagodziński et al. 2012; Jagodzinski et al. 2009). Jagodzinski et al. (2009) tried to compare tsunami deposits, beach sediments, and pre-tsunami soils in Thailand and Japan. In Thailand, the difference between tsunami deposits and beach sediments and soils was reflected in differences in the respective proportions of mica and tourmaline. These differences were attributed to the mode of sediment transport and deposition with mica, due to its low density, being more abundant in the top-most part of the tsunami deposit.

Scanning electron microscopy (SEM) of predominantly quartz grains has also been used to establish the source material of tsunami deposits (e.g., Bruzzi and Prone 2000; Costa et al. 2012b; Costa et al. 2009; Dahanayake and Kulasena 2008). Anisotropy of magnetic susceptibility has also been used to provenance studies of tsunamigenic deposits (e.g., Font et al. 2010; Sugawara et al. 2008; Wassmer et al. 2010), but the application of this technique is still in its infancy and always requires that the data produced is normalized in respect of the grain-size distribution. One example of these studies was conducted by Font et al. (2010) in Boca do Rio (Portugal) where the magnetic data showed a dominance of paramagnetic minerals (quartz) mixed with lesser amount of ferromagnetic minerals, namely, titanomagnetite and titanohematite, both of a detrital origin and reworked from the underlying sedimentary units. Statistical analyses pointed to a scenario where the energy released by the tsunami wave was strong enough to overtop and erode sand from the littoral dune and mixed it with reworked materials from underlying layers from at least 1 m in depth (see Table 1).

Sedimentary Structures

In terms of sedimentological structures, an erosive/sharp/abrupt basal contact is a common feature and is symptomatic of the energy involved in the emplacement of tsunami deposits. However, this criterion was also recognized in storm deposits (e.g., Morton et al. 2007). The sharp/abrupt/erosive contact of tsunamigenic layers was firstly described by Dawson et al. (1988). Bondevik et al. (1997), analyzing evidences laid down by the Storegga tsunami in Norway, detected that the tsunami deposit rests on an erosional unconformity which in cases has removed more than 1 m of underlying sediment. Moreover, Nanayama et al. (2000) observed deposits that resulted from the 1993 Hokkaido-Nansei-Oki (Japan) tsunami and identified distinctive sharp erosional bases in the tsunamigenic unit. Gelfenbaum and Jaffe (2003) analyzed the erosion and sedimentation associated with the 1998 Papua New Guinea

tsunami and observed that the beach face and berm showed no evidence of deposition from the tsunami. However, on the berm, exposed roots and scour at the base of some palm trees indicated erosion of approx. 20–30 cm of backbeach sand, and they observed that only erosional signatures had been left by this tsunami to the landward side of the berm, up to about 50 m from the shoreline. In Thailand, Szczucinski et al.(2006), Hori et al.(2007), and Fujino et al.(2009) observed an erosive sharp basal contact between the tsunamigenic and the underlying layers. Choowong et al.(2009) also noted that during the same event, erosion and deposition occurred mainly during two periods of inflow and that the return flow was mainly erosive. Paris et al.(2009) described erosion associated with the 2004 Indian Ocean tsunami in Banda Aceh (Indonesia) that extended up to 500 m inland. These authors quantified the overall coastal retreat from Lampuuk to Leupung as of the order of 60 m (*ca.* 550,000 m²) and locally in excess of 150 m. The erosional impact of tsunamis is still controversial, not only the recognition of associated patterns in the sedimentary record, other than the erosive base and quantification of the amount of sediment removed but also the mechanisms and processes associated and responsible for the erosional/depositional balance during a tsunami. In fact, Bahlburg and Spiske (2011) analyzing the sedimentary record of the February 2010 tsunami at Isla Mocha (Chile) observed that the tsunamigenic unit was produced essentially (i.e., >90 %) by the backflow. These authors suggest that due to the lack of sedimentary structures, many previous studies of modern tsunami sediments assumed that most of the detritus were deposited during inflow and an uncritical use of this assumption may lead to erroneous interpretations of palaeotsunami magnitudes and sedimentary processes if unknowingly applied to backflow deposits. Typically, tsunami deposits present sediment size that can vary from mud to boulders, and, in many cases, grain-size variation in tsunami deposits is controlled by the size of sediment available for transport, rather than by flow capacity or direction (Bourgeois 2009).



Tsunami Sedimentology, Fig. 4 Image of a mud drape in a tsunamigenic deposit in Martinhal (*left image*) and image of a rip-up mud ball in a tsunamigenic deposit from Boca do Rio (*right image*). Photos by C. Freitas

The detection of sedimentary structures is limited by sampling methods because coring (which is frequently used), in contrast to trench excavation, is in general destructive. Sedimentary structures are also difficult to identify in tsunami deposits due to the common deposition as massive deposit (e.g., Dahanayake and Kulaseena 2008). However, there have been several tsunami deposits where sedimentary structures include *laminae* (e.g., Bondevik et al. 2005), rip-up clasts (Fig. 4 and, e.g., Gelfenbaum and Jaffe 2003; Goff et al. 2004; Hindson and Andrade 1999; Morton et al. 2007), and cross-stratification (e.g., Choowong et al. 2008). Muddy *laminae* or organic layers can represent evidence for multiple waves of the tsunami wave train (e.g., Bondevik et al. 1997). Furthermore, loading structures at the base of the deposit have been reported in literature (e.g., Costa 2006; Minoura et al. 1996).

Another feature observed in many tsunamigenic deposits worldwide is the enrichment in bioclasts or shells (many of them broken) when compared with the under- and overlying layers (e.g., Costa 2006), and in cases platy or prolate shell fragments occur aligned suggesting a ghostly lamination (e.g., Dawson et al. 1995). For example, moreover, Donato et al. (2008) showed that shell features could be used as useful indicators of tsunamigenic deposit due to their vertical and lateral extent, to the allochthonous mixing of articulated bivalve species (e.g., lagoonal and

nearshore) out of life position, and to the high amount of fragmented valves, with angular breaks and stress fractures. The authors suggested that the taphonomic uniqueness of tsunami deposits should be considered as a valid tool for tsunamigenic recognition in the geological record.

The sedimentological fingerprint of currents associated with tsunami events has also been observed in the form of parallel lamination, cross-lamination, convolutions, and ripple marks (e.g., Shiki et al. 2008). Moreover, Morton et al. (2007) detected palaeocurrent indicators in tsunami deposits indicating seaward return flow. In deposits laid down by the 2004 Indian Ocean tsunami, in Thailand, Choowong et al. (2008) observed capping bedforms and parallel *laminae*, cross-lamination, rip-up mud, and sand clasts. The authors also observed normal grading but some reverse grading was locally recognized. According to Choowong et al. (2008), reverse grading in tsunami deposits indicates a very high grain concentration within the tsunami flow and was possibly formed at the initial stages of inundation in shallow water. Cross-bedding was seen as restricted to return flow sediments (Nanayama et al. 2000). Individual deposits are generally well sorted (many are massive) and characterized by sets of fining-upward sediment sequences that were interpreted by Shi and Smith (2003) as indicative of deposition by individual tsunami waves. Dawson and Smith (2000)

characterized a tsunami sequence in Scotland by several fining-upward sequences indicative of a series of tsunami waves and episodes of backwash. Furthermore, run-up and return flow deposits were also differentiated by Dawson et al. (1996a) or Nanayama et al. (2000). In the case of the Indian Ocean tsunami, Paris et al. (2007) observed a landward sequence thinning, fining, and sorting. Normally graded couplets or triplets of layers were used to identify the run-up of each wave. The topmost layers, interpreted as the backwash deposition, describe a seaward sequence of decreasing mean grain size.

The time lag separating tsunami wave trains is occasionally marked between different subunits by the presence of mud drapes (Fig. 4). Moreover, Fujiwara and Kamataki (2007) observed the presence of a vertical stack of many coarse-grained sub-layers separated by mud drapes interpreted as due to incremental deposition from multiple sediment flows separated by flow velocity stagnation stages and concluded that it is unlikely that the mud drapes were deposited by short-period storm waves.

Palaeontology

More recently, different techniques have been explored to identify the sedimentary structures in tsunami deposits. An example of this is ground-penetrating radar, used by Switzer et al. (2006) to survey the erosional contact between an event layer and the under- and overlying units. Koster (2012) also used ground-penetrating radar in combination with electrical resistivity tomography measurements and sedimentology for tsunamiite recognition in Greece and Spain. According to these authors, ground-penetrating radar data indicated unconformable thicknesses of tsunamigenic beddings, channel-like structures (backwash deposits), and to some extent basal erosion, as well as abrasion-scours in various places, and boulder accumulation inside the deposits (see Table 1).

Macrofossils and microfossils have been used to identify and interpret sedimentary units as tsunamigenic. To date, the use of palaeontological characteristics to recognize

tsunami deposits has focused on diatoms, foraminifera, ostracods, nannoplankton, pollen, mollusks, and plant fragments. Typically, the palaeontological signature is characterized by marked changes in the population indicating the increase in abundance of marine to brackish fossils and/or the high energy of the event (e.g., presence of broken shells, plant fragments, parts of twigs and roots, etc.). Generally, diatom assemblages in tsunami deposits are chaotic (mixture of freshwater and brackish-marine species), because tsunami crosses coastal and inland areas eroding, transporting, and redepositing freshwater taxa (e.g., Dawson et al. 1996; Sawai et al. 2009).

Foraminiferal content is also a common micropalaeontological proxy that has been used in tsunami sediment provenance studies (e.g., Hindson et al. 1996; Kortekaas and Dawson 2007; Mamo et al. 2009). Mamo et al. (2009) summarized the many procedures, characteristics, and limitations associated with foraminiferal assemblages and their use in the recognition of tsunami deposits. Characteristics such as changes in assemblage composition, for example, marine shelf species within a lagoon or brackish environment, changes in test size or in juvenile to adult ratios, a shift in population numbers, or a change in the taphonomic character of the tests can be used to recognize tsunami deposits. Given that the exact composition of an assemblage varies from location to location, it is impossible to expect to see a specific diagnostic specie(s) or assemblage in association with tsunami-deposited sediments.

Several studies were conducted in tsunami deposits with the aim of identifying a distinctive geochemical signature (e.g., Chagué-Goff 2010). Usually geochemical features of tsunami deposits simply indicate the presence of saltwater inundation and thus do not provide information on the specific type of inundation. However, increases in the concentration of chemical elements of marine origin or elements indicative of coarser-sized sediments have been recognized in the past as a proxy in the study of extreme marine inundations. This is often observed without a distinctive change in lithostratigraphy.

The preservation of tsunami deposit is a fundamental factor in any sedimentological study focusing in recurrence intervals of such events. In fact, Szczuciński (2012) conducted five yearly surveys after the 2004 Indian Ocean tsunami and concluded that the post-tsunami recovery of coastal zones was generally in the order of a few months to a few years. The same author observed that almost all the near-surface structures of the tsunami deposits were removed with time (i.e., after at least one rainy season). Tsunami deposits thinner than 10 cm usually acquired a massive appearance after 1 or 2 years; the only remnants of the primary structures, for instance, fining upward, were removed. This was attributed to bioturbation by vigorous growing root systems and burrowing animals like crabs and rodents.

According to Yawsangratt et al. (2012), micropalaeontological evidences (i.e., carbonate foraminifera) may be subjected to significant dissolution 4.5 years after tsunami emplacement; again, this postdepositional disturbance is not exclusive of tsunami deposits, and rapid intrasediment dissolution or downwearing of carbonate foraminifera tests, Ostracoda valves, or diatom frustules is a common drawback micropalaeontologists working in Holocene sediments of various facies are used to and aware of and not exclusive of high-energy events of abrupt marine inundation.

Age Estimation Methods

It is important to accurately establish the dates of deposition of a specific event. However, many of the dating techniques currently employed in tsunami deposit studies can be applied only to restricted spans of time, and each method has its own distinctive set of problems which lead to uncertainties in the interpretation. The techniques used (and explained below) are the more commonly used in the establishment of an age for the deposition of sediments laid down during tsunami events. However, dating of abrupt marine invasions is, in most cases, problematic. Results tend to be difficult to interpret, and in many cases the dates obtained are not accurate or even in stratigraphical order. The approach commonly

used is, to date, not only the tsunamigenic layer but also the under- and overlying units. The combination of these results allows an accurate establishment of the age of deposition.

Optically stimulated luminescence dating (a development of thermoluminescence) is used for dating sediments, where exposure to light prior to burial has been limited. This technique measures the luminescence emitted from the most light-sensitive electron traps in minerals (e.g., quartz and feldspars). Luminescence dating methods have previously been applied to tsunami deposits using thermoluminescence, infrared stimulated luminescence, and optically stimulated luminescence (OSL; e.g., Kortekaas and Dawson 2007).

^{210}Pb is an unstable isotope present in the atmosphere and is produced by the decay of ^{222}Rn which is part of the U series – decay chain. This isotope is absorbed by the sediments where it decays to stable ^{206}Pb over an interval of circa 150 years. By measuring the ratio between ^{210}Pb and ^{206}Pb in a column of sediment in relation to depth, and assuming that the atmospheric flux of ^{210}Pb has remained constant, the time that has elapsed since the lead was deposited can be determined. The sedimentation rate can then be established.

This method has however a problem related with the differentiation of supported ^{210}Pb (resulting from the Uranium decay) and the unsupported ^{210}Pb (produced in the atmosphere). The absolute determination of ^{210}Pb is obtained by measuring the alpha activities of the isotope. It is common to use internal yield tracers in isotope dilution to achieve more accurate results compared with gross alpha counting. Isotope dilution alpha spectrometry requires the application of chemical procedures that remove interfering species and provide thin alpha sources with well-resolved alpha peaks.

^{210}Pb can be directly measured with gamma spectrometry, but the more sensitive alpha spectrometry method (by measuring it indirectly via its granddaughter ^{210}Po) is preferred. The advantage of alpha spectrometry is that it requires less sediment and allows more samples to be counted simultaneously.

^{137}Cs is an artificially generated radioactive nuclide (half-life of 30 years) that has only been produced in significant quantities as a result of thermonuclear weapons testing which began in 1945. The temporal patterns of cesium-137 input can be characterized by: (1) detectable cesium-137 began in 1954; (2) the first peak appeared in 1958/1959; (3) the second peak occurred at 1962–1964; and (4) the termination of cesium-137 input around mid-1980s. Some areas had additional input in 1986 after the Chernobyl incident. This annual variation has been successfully used to determine sediment accumulation rates in a wide variety of depositional environments including reservoirs, lakes, wetlands, coast areas, and flood plains.

Identification of cesium-137 loss and gain for each sampled location in the study field is made by comparison of the measurement done using gamma spectrometry from the reference sites with those from the study field, allowing the establishment of a sedimentation rate.

The most common age estimation method applied to palaeotsunami research is radiocarbon dating. ^{14}C is continually being produced in the upper atmosphere, becomes stored in various global reservoirs (atmosphere, biosphere, and hydrosphere), and is absorbed by living organisms. All living organisms absorb carbon dioxide during tissue building in a ratio that is broadly in equilibrium with atmospheric carbon dioxide. After the death of the organisms, the ^{14}C decays to ^{14}N . The activity of ^{14}C in the atmosphere is known (15 disintegrations per minute per gram). The half-life of ^{14}C was originally calculated at 5568 ± 30 years, but subsequently this has been more accurately determined as 5730 ± 40 years. However, the value that is used by convention is the former due to the large number of dates that had been produced prior to the new measurement. The logarithmic decay curve for radiocarbon has been established, and calculations can be made using this curve (up to approx. 50,000 years, which is approximately eight times the half-life of ^{14}C).

Two approaches are used to measure the residual ^{14}C activity in a sample: the conventional radiocarbon dating or the accelerator mass spectrometry. The first involves the detection and

counting of β emissions from ^{14}C atoms over a period of time in order to determine the rate of emissions and hence the activity of the sample. In the calculation of radiocarbon dates obtained by conventional methods, laboratories compare sample activities to a modern reference standard. The second uses particle accelerators as mass spectrometers (AMS) to count the actual number of ^{14}C atoms as opposed to their decay products in a sample of material. Comparing the two methodologies, one can conclude that AMS allows the measurements of smaller samples and is quicker and more reliable.

The sources of error in ^{14}C dating are the temporal variations in ^{14}C production, the isotopic fractionation, the circulation of marine carbon, the contamination of the samples (especially in shells – often used in palaeotsunami research), and the biogeochemistry of lake sediments.

Summary

In summary, it is important to note that the criteria to recognize tsunami deposits (see Table 1) are still, at the present state of knowledge, ambiguous. Together, the identification of sedimentary structures, the establishment of source material, the micropalaeontological analysis, the geochemical characteristics, and the study of geomorphological imprints in coastal landscape facilitate the identification of tsunami deposits, even if erosion and preservation are constrained by several facts. In fact, the contrast/peculiarity of the tsunami layers, especially when compared with under- and overlying layers, provides in many cases the conclusive evidence for the recognition of such deposits. Data is commonly obtained through the use of sedimentological techniques – some have been widely used (e.g., textural and geomorphological) while others have been scarcely applied (e.g., microtextural analysis and heavy mineral assemblages); both have been used in modern tsunami sediments as well as in palaeotsunamis. It is possible, through the use of diverse sedimentological proxies, to obtain information about the presence or absence of tsunami indicators, to establish their likely source, or to collect valuable information about tsunami run-up, backwash, or wave penetration inland. Recent events in Sumatra and

Japan have been used to further develop the application and definition of sedimentary criteria to be used in the identification of tsunami deposits. However, in the study of palaeotsunamis, further key questions that remain (e.g., understanding mechanisms of inundation and deposition for each specific location, preservation of sedimentary structures, and palaeontological evidence) still need to be addressed in future studies in order to contribute to the development of more detailed and rigorous criteria that can further contribute to the accuracy of tsunami hazard maps.

Future Directions

Constraints on our present-day ability to separate and identify the geological evidence of extreme marine inundations in the sedimentary record (e.g., associated with preservation of deposit, existence of lithological and textural contrasts, etc.) need further research to establish return periods more confidently.

Reexamination and more detailed provenance studies should be conducted with the purpose of establishing tsunami- and storm-deposit sedimentary sources and, through this, to improve the understanding of the physical repercussions into the sedimentary processes characteristic of extreme marine inundations. For instance, the backwash process, and its influence in the arrangement of tsunamigenic units, is at the present state of knowledge scarcely understood. To conduct these studies, the widespread use of microtextural and heavy mineral analysis on samples obtained from both observed and fossil deposits is recommended; revisiting classic reference locations and deposits, resampling and reinterpreting these “classical” deposits, and incorporating the developments in understanding backwash signatures from recent events may result in significant advances in the comprehension of the geological record. Furthermore, the increase in the number of research sites will facilitate generalization of conclusions and perception on which proxies or features may be used, independently of regional to local conditions, thus diminishing the influence of site-specific

constraints. These approaches will provide more tools and data to comprehend depositional and erosional processes involved in tsunami and storm sediment flow and transport.

In some cases, where tsunami or storms present significant sedimentary erosive patterns (volume of sediment remobilized exceeding the total volume of the sediment unit inland), it would be challenging and innovative to focus on the sediment deposited at the offshore area, beyond the closure depth, where some of these deposits are most probably immobilized and in a state of “nonequilibrium” with the prevailing oceanographic conditions; if that section of the inner shelf is accreting, the potential of a short core containing record of multiple events is high.

Further studies should address the potential sedimentary differences between tsunami deposits caused by different origins (i.e., earthquake and landslide) with the ambitious aim of increasing our ability to distinguish sedimentary columns containing multiple events generated by the same source (even if separated in space) from sections containing signatures of multiple-sourced events, both cases being of different significance in geodynamic and hazard terms. Furthermore, the relationship between distant and local tsunami origins will most probably (at least, at a regional scale) be reflected in the tsunami deposit. The application and development of more accurate age estimation techniques (direct or indirect) will also facilitate the establishment of return periods of extreme marine inundations with obvious positive consequences for the study and assessment of hazard/risk for any specific coastal area. It is essential that future research contributes to the increase of the accuracy of the sedimentological criteria to recognize and differentiate sedimentary products laid down by extreme marine inundations.

Bibliography

Primary Literature

- Andrade C (1992) Tsunami generated forms in the Algarve Barrier Islands (South Portugal). *Sci Tsunami Haz* 10(1):21–34
- Babu N, Suresh Babu DS, Mohan Das PN (2007) Impact of tsunami on texture and mineralogy of a major placer

- deposit in southwest coast of India. *Environ Geol* 52:71–80
- Bahlburg H, Spiske M (2011) Sedimentology of tsunami inflow and backflow deposits: key differences revealed in a modern example. *Sedimentology*. doi:10.1111/j.1365-3091.2011.01295.x
- Bahlburg H, Weiss R (2006) Sedimentology of the December 26, 2004, Sumatra tsunami deposits in eastern India (Tamil Nadu) and Kenya. *Int J Earth Sci (Geol Rundsch)*. doi:10.1007/s00531-006-0148-9, 1–15.
- Bondevik S, Svendsen JJ, Mangerud J (1997) Tsunami sedimentary facies deposited by the Storegga tsunami in shallow marine basins and coastal lakes, western Norway. *Sedimentology* 44:1115–1131
- Bondevik S, Mangerud J, Dawson S, Dawson A, Lohne Å (2005) Evidence for three North Sea tsunamis at the Shetland Islands between 8000 and 1500 years ago. *Quat Sci Rev* 24:1757–1775
- Bruzzi C, Prone A (2000) A method of sedimentological identification of storm and tsunami deposits: exoscopic analysis, preliminary results. *Quaternaire* 11(3–4):167–177
- Chagué-Goff C (2010) Chemical signatures of palaeotsunamis: a forgotten proxy? *Mar Geol* 271:67–71
- Chagué-Goff C, Dawson S, Goff JR, Zachariassen J, Berlyman KR, Garnett DL, Waldron HM, Mildenhall DC (2002) A tsunami (ca. 6300 years BP) and other Holocene environmental changes, northern Hawke's Bay, New Zealand. *Sediment Geol* 150:89–102
- Chagué-Goff C, Schneider J-L, Goff JR, Dominey-Howes D, Strotz L (2011) Expanding the proxy toolkit to help identify past events Lessons from the 2004 Indian Ocean Tsunami and the 2009 South Pacific Tsunami. *Earth Sci Rev* 107:107–122
- Choowong M, Murakoshi N, Hisada K-I, Charusiri P, Charoentitirat T, Chutakositkanon V, Jankaew K, Kanjanapayont P, Phantuwongraj S (2008) 2004 Indian Ocean tsunami inflow and outflow at Phuket, Thailand. *Mar Geol* 248:179–192
- Choowong M, Phantuwongraj S, Charoentitirat T, Chutakositkanon V, Yumuang S, Charusiri P (2009) Beach recovery after 2004 Indian Ocean tsunami from Phang-nga, Thailand. *Geomorphology* 104:134–142
- Clague JJ, Bobrowsky PT, Hutchinson I (2000) A review of geological records of large tsunamis at Vancouver Island, British Columbia, and implications for hazard. *Quaternary Science Reviews* 19:849–863
- Costa PJM (2006) Geological recognition of abrupt marine invasions in two coastal areas of Portugal, MPhil thesis, p. 139
- Costa PJM, Andrade C, Freitas MC, Oliveira MA, Jouanneau JM (2009) Preliminary results of exoscopic analysis of quartz grains deposited by a Palaeotsunami in Salgados Lowland (Algarve, Portugal). *J Coast Res* 56:39–43
- Costa PJM (2012), Sedimentological signatures of extreme marine inundations, PhD thesis, 245 pp, Universidade de Lisboa, Portugal.
- Costa PJM, Andrade C, Dawson AG, Mahaney WC, Freitas MC, Paris R, Taborda R (2012a) Microtextural characteristics of quartz grains transported and deposited by tsunamis and storms. *Sediment Geol* 275:55–69
- Costa PJM, Andrade C, Freitas MC, Oliveira MA, Lopes V, Dawson AG, Moreno J, Fatela F, Jouanneau JM (2012b) A tsunami record in the sedimentary archive of the central Algarve coast, Portugal: characterizing sediment, reconstructing sources and inundation paths. *Holocene* 22:899–914
- Costa PJM, Andrade C, Cascalho J, Dawson AG, Freitas MC, Paris R, Dawson S (2015) Heavy mineral assemblages of onshore palaeotsunami sediments. *The Holocene* 25(5):795–809
- Dahanayake K, Kulaseena N (2008) Recognition of diagnostic criteria for recent- and paleo-tsunami sediments from Sri Lanka. *Mar Geol* 254:180–186
- Dawson AG (1994) Geomorphological effects of tsunami run-up and backwash. *Geomorphology* 10:83–94
- Dawson AG (1999) Linking tsunami deposits, submarine slides and offshore earthquakes. *Quat Int* 60:119–126
- Dawson S, Smith DE (2000) The sedimentology of Middle Holocene tsunami facies in northern Sutherland, Scotland, UK. *Mar Geol* 170:69–79
- Dawson AG, Stewart I (2007) Tsunami deposits in the geological record. *Sediment Geol* SEDGEO-03769:18
- Dawson AG, Long D, Smith DE (1988) The storegga slides: evidence from eastern Scotland for a possible tsunami. *Mar Geol* 82:271–276
- Dawson AG, Hindson R, Andrade C, Freitas C, Parish R, Bateman R (1995) Tsunami sedimentation associated with the Lisbon earthquake of 1 November AD 1755: Boca do Rio, Algarve, Portugal. *The Holocene* 5(2):209–215
- Dawson AG, Shi S, Dawson S, Takahashi T, Shuto N (1996a) Coastal sedimentation associated with the June 2nd and 3rd, 1994 tsunami in Rajegwesi, Java. *Quat Sci Rev* 15:901–912
- Dawson S, Smith DE, Ruffman A, Shi S (1996b) The diatom biostratigraphy of tsunami sediments: examples from recent and middle holocene events. *Phys Chem Earth* 21:87–92
- Donato SV, Reinhardt EG, Boyce JJ, Rothaus R, Vosmer T (2008) Identifying tsunami deposits using bivalve shell taphonomy. *Geology* 36(3):199–202
- Font E, Nascimento C, Omira R, Baptista MA, Silva PF (2010) Identification of tsunami-induced deposits using numerical modeling and rock magnetism techniques: a study case of the 1755 Lisbon tsunami in Algarve, Portugal. *Phys Earth Planet In* 182:187–198
- Fujino S, Naruse H, Matsumoto D, Jarupongsakul T, Sphawajruksakul A, Sakakura N (2009) Stratigraphic evidence for pre-2004 tsunamis in southwestern Thailand. *Mar Geol* 262:25–28

- Gelfenbaum G, Jaffe B (2003) Erosion and Sedimentation from the 17 July, 1998, Papua New Guinea Tsunami. *Pure Appl Geophys* 160:1969–1999
- Goff J, McFadgen BG, Chagué-Goff C (2004) Sedimentary differences between the 2002 Easter storm and the 15th-century Okoropunga tsunami, southeastern North Island, New Zealand. *Mar Geol* 204:235–250
- Goff J, McFadgen B, Wells A, Hicks M (2008) Seismic signals in coastal dune systems. *Earth Sci Rev* 89:73–77
- Goff J, Pearce S, Nichol SL, Chagué-Goff C, Horrocks M, Strotz L (2010a) Multi-proxy records of regionally-sourced tsunamis, New Zealand. *Geomorphology* 118:369–382
- Goff J, Weiss R, Courtney C, Dominey-Howes D (2010b) Testing the hypothesis for tsunami boulder deposition from suspension. *Mar Geol* 277:73–77
- Goto K, Kawana T, Imamura F (2010a) Historical and geological evidence of boulders deposited by tsunamis, southern Ryukyu Islands, Japan. *Earth Sci Rev* 102:77–99
- Goto K, Okada K, Imamura F (2010b) Numerical analysis of boulder transport by the 2004 Indian Ocean tsunami at Pakarang Cape, Thailand. *Mar Geol* 268:97–105
- Hindson RA, Andrade C (1999) Sedimentation and hydrodynamic processes associated with the tsunami generated by the 1755 Lisbon earthquake. *Quat Int* 56:27–38
- Hindson RA, Andrade C, Dawson AG (1996) Sedimentary processes associated with the tsunami generated by the 1755 Lisbon earthquake on the Algarve coast, Portugal. *Phys Chem Earth* 21:57–63
- Hori K, Kuzumoto R, Hirouchi D, Umitsu M, Janjirawuttikul N, Patanakanog B (2007) Horizontal and vertical variation of 2004 Indian tsunami deposits: an example of two transects along the western coast of Thailand. *Mar Geol* 239:163–172
- Jagodzinski R, Sternal B, Szczucinski W, Lorenc S (2009) Heavy minerals in 2004 tsunami deposits on Kho Khao Island, Thailand. *Pol. J Environ Studies* 18:103–110
- Jagodziński R, Sternal B, Szczuciński W, Chagué-Goff C, Sugawara D (2012) Heavy minerals in the 2011 Tohoku-oki tsunami deposits – insights into sediment sources and hydrodynamics. *Sediment Geol* 282:57–64
- Kortekaas S, Dawson AG (2007) Distinguishing tsunami and storm deposits: an example from Martinhal, SW Portugal. *Sediment Geol* 200(3–4):208–221
- Koster B (2012) Ground penetrating radar (GPR) investigations on tsunami-migenic deposits – Examples from southern Spain and Greece. *Quat Int* 279–280:254
- Koster B (2014). Modern approaches in palaeotsunami research. PhD thesis, 146 pp., RWTH University, Germany.
- Mamo B, Strotz L, Dominey-Howes D (2009) Tsunami sediments and their foraminiferal assemblages. *Earth Sci Rev* 96:263–278
- Minoura K, Gusiakov VG, Kurbatov A, Takeuti S, Svendsen JJ, Bondevik S, Oda T (1996) Tsunami sedimentation associated with the 1923 Kamchatka earthquake. *Sediment Geol* 106:145–154
- Moore AL, McAdoo BG, Ruffman A (2007) Landward fining from multiple sources in a sand sheet deposited by the 1929 Grand Banks tsunami, Newfoundland. *Sediment Geol* 200:336–346
- Morton RA, Gelfenbaum G, Jaffe BE (2007) Physical criteria for distinguishing sandy tsunami and storm deposits using modern examples. *Sediment Geol* 200:184–207
- Morton RA, Goff JR, Nichol SL (2008) Hydrodynamic implications of textural trends in sand deposits of the 2004 tsunami in Sri Lanka. *Sediment Geol.* doi:10.1016/j.sedgeo.2008.03.008, 40
- Nanayama F, Shigeno K, Satake K, Shimokawa K, Koitabashi S, Miyasaka S, Ishii M (2000) Sedimentary differences between the 1993 Hokkaido-nansei-oki tsunami and the 1959 Miyakojima typhoon at Taisei, southwestern Hokkaido, northern Japan. *Sediment Geol* 135:255–264
- Nanayama F, Furukawa R, Shigeno K, Makino A, Soeda Y, Igarashi Y (2007) Nine unusually large tsunami deposits from the past 4000 years at Kiritappu marsh along the southern Kuril Trench. *Sediment Geol* 200:275–294
- Nandasena NAK, Paris R, Tanaka N (2011) Reassessment of hydrodynamic equations: minimum flow velocity to initiate boulder transport by high energy events (storms, tsunamis). *Mar Geol* 281:70–84
- Narayana AC, Tataavarti R, Shinu N, Subeer A (2007) Tsunami of December 26, 2004 on the southwest coast of India: post-tsunami geomorphic and sediment characteristics. *Mar Geol* 242:155–168
- Nott J (1997) Extremely high-energy wave deposits inside the Great Barrier Reef, Australia: determining the cause – tsunami or tropical cyclone. *Mar Geol* 141:193–207
- Paris R, Lavigne F, Wassmer P, Sartohadi J (2007) Coastal sedimentation associated with the December 26, 2004 tsunami in Lhok Nga, west Banda Aceh (Sumatra, Indonesia). *Mar Geol* 238:93–106
- Paris R, Wassmer P, Sartohadi J, Lavigne F, Barthomeuf B, Desgages E, Grancher D, Baumert P, Vautier F, Brunstein D, Gomez C (2009) Tsunamis as geomorphic crises: lessons from the December 26, 2004 tsunami in Lhok Nga, West Banda Aceh (Sumatra, Indonesia). *Geomorphology* 104:59–72
- Sawai Y, Jankaew K, Martin ME, Prendergast A, Choowong M, Charoentitirat T (2009) Diatom assemblages in tsunami deposits associated with the 2004 Indian Ocean tsunami at Phra Thong Island, Thailand. *Mar Micropaleontol* 73:70–79
- Scheffers A, Kelletat D (2005) Tsunami relics on the coastal landscape west of Lisbon, Portugal. *Sci Tsunami Haz* 23(1):3–16
- Scheffers A, Scheffers S (2007) Tsunami deposits on the coastline of west Crete (Greece). *Earth Planet Sci Lett.* doi:10.1016/j.epsl.2007.05.041, 32

- Scheffers A, Shiki T, Tsuji Y, Yamazaki T, Minoura K (2008) Tsunami boulder deposits, Tsunamiites. Elsevier, Amsterdam, pp 299–317
- Shennan I, Long AJ, Rutherford MM, Green FM, Innes JB, Lloyd JM, Zong Y, Walker KJ (1996) Tidal marsh stratigraphy, sea-level change and large earthquakes, i: a 5000 year record in Washington, U.S.A. *Quat Sci Rev* 15:1023–1059
- Shi S, Smith DE (2003) Coastal tsunami geomorphological impacts and sedimentation processes: case studies of modern and prehistorical events. In: International conference on estuaries and coasts, 9–11 Nov 2003, Hangzhou
- Shiki T, Tachibana T, Fujiwara O, Goto K, Nanayama F, Yamazaki T, Shiki T, Tsuji Y, Yamazaki T, Minoura K (2008) Characteristic features of Tsunamiites, Tsunamiites. Elsevier, Amsterdam, pp 319–340
- Srinivasulu S, Thangadurai N, Switzer AD, Mohan VR, Ayyamperumal T (2007) Erosion and sedimentation in Kalpakkam (N Tamil Nadu, India) from the 26th December 2004 tsunami. *Mar Geol* 240:65–75
- Sugawara D, Minoura K, Imamura F, Shiki T, Tsuji Y, Yamazaki T, Minoura K (2008) Tsunamis and tsunami sedimentology, Tsunamiites. Elsevier, Amsterdam, pp 9–49
- Switzer AD, Pucillo K, Haredy RA, Jones BG, Bryant EA (2005) Sea-level, storms or tsunami; enigmatic sand sheet deposits in sheltered coastal embayment from southeastern New South Wales Australia. *J Coast Res* 21:655–663
- Switzer AD, Bristow CS, Jones BG (2006) Investigation of large-scale washover of a small barrier system on the southeast Australian coast using ground penetrating radar. *Sediment Geol* 183:145–156
- Switzer AD, Jones BG (2008) Large-scale washover sedimentation in a freshwater lagoon from the southeast Australian coast: tsunami or exceptionally large storm? *The Holocene*, 18:787–803
- Szczuciński W (2012) The post-depositional changes of the onshore 2004 tsunami deposits on the Andaman Sea coast of Thailand. *Nat Hazards* 60:115–133
- Szczucinski W, Chaimanee N, Niedzielski P, Rachlewicz G, Saisuttichai D, Tepsuwan T, Lorenc S, Siepak J (2006) Environmental and geological impacts of the 26 December 2004 Tsunami in coastal zone of Thailand – overview of short and long-term effects. *Pol J Environ Stud* 15(5):793–810
- Wassmer P, Schneider J-L, Fonfrige A-V, Lavigne F, Paris R, Gomez C (2010) Use of anisotropy of magnetic susceptibility (AMS) in the study of tsunami deposits: application to the 2004 deposits on the eastern coast of Banda Aceh, North Sumatra, Indonesia. *Mar Geol* 275:255–272
- Yawsangratt S, Szczucinski W, Chaimanee N, Chatprasert S, Majewski W, Lorenc S (2012) Evidence of probable paleotsunami deposits on Kho Khao Island, Phang Nga Province, Thailand. *Nat Hazards*. doi:10.1007/s11069-011-9729-4
- ## Books and Reviews
- Atwater BF, Moore AL (1992) A tsunami about 1000 years ago in Puget Sound, Washington. *Science* 258:1614–1617
- Bourgeois J (2009) *Geologic records and effects of tsunamis: in the sea*. Harvard University Press, Cambridge
- Bryant E, Nott J (2001) Geological indicators of large tsunami in Australia. *Nat Hazards* 24:231–249
- Bryant E, Young R, Price D (1992) Evidence of tsunami sedimentation on the southeastern coast of Australia. *J Geol* 100:753–765
- Chandrasekar N (2005) Tsunami of 26th December 2004: observation on inundation, sedimentation and geomorphology of Kanyakumari coast, South India. In: Satake PA (ed) 22nd Intl. Tsunami Symposium. Xania, pp. 49–56
- Costa PJM, Andrade C, Freitas MC, Oliveira MA, Silva CMD, Omira R, Taborda R, Baptista MA, Dawson AG (2011) Boulder deposition during major tsunami events. *Earth Surf Processes Landf* 36:2054–2068
- Dawson S (2007) Diatom biostratigraphy of tsunami deposits: examples from the 1998 Papua New Guinea tsunami. *Sediment Geol*. doi:10.1016/j.sedgeo.2007.01.011, 8
- Dominey-Howes D (2007) Geological and historical records of tsunami in Australia. *Mar Geol* 239:99–123
- Fujiwara O, Kamataki T (2007) Identification of tsunami deposits considering the tsunami waveform: an example of subaqueous tsunami deposits in Holocene shallow bay on southern Boso Peninsula, Central Japan. *Sediment Geol* 200:295–313
- Hemphill-Haley E (1996) Diatoms as an aid in identifying late-Holocene tsunami deposits. *The Holocene* 6:439–448
- Higman B, Bourgeois J (2008) Deposits of the 1992 Nicaragua tsunami. In: Shiki T, Tsuji Y, Yamazaki T, Minoura K (eds) *Tsunamiites – features and implications*. Elsevier, Amsterdam, pp 81–103
- Higman B, Jaffe B (2005) A comparison of grading in deposits from five tsunamis: does tsunami wave duration affect grading patterns?, *Eos (Transactions, American Geophysical Union)* 86, Abstract T11A-0362
- Imamura F, Lee HJ, Takahashi T, Shuto N (1997) The highest run-up of the 1993 Hokkaido Nansei-Oki earthquake tsunami. *IAMAS – IAPSO Joint Assemblies*, Melbourne
- Lakshmi CSV, Srinivasan P, Murthy SGN, Trivedi D, Nair RR (2010) Granularity and textural analysis as a proxy for extreme wave events in southeast coast of India. *J Earth Syst Sci* 119:297–305
- Matsumoto D, Naruse H, Fujino S, Surphawajraksakul A, Jarupongsakul T, Sakakura N, Murayama M (2008) Truncated flame structures within a deposit of the Indian Ocean Tsunami: evidence of syndimentary deformation. *Sedimentology* 55:1559–1570
- Paris R, Pérez Torrado FJ, Carracedo JC (2005) Massive flank failures and tsunamis in the Canary Islands: past,

- present, future. *Zeitschrift für Geomorphologie* 140:37–54
- Paris R, Naylor LA, Stephenson WJ (2011) Boulders as a signature of storms on rock coasts. *Mar Geol* 283:1–11
- Phantuwongraj S, Choowong M (2011) Tsunamis versus storm deposits from Thailand. *Nat Hazards* 63:31–50
- Sato H, Shimamoto T, Tsutsumi A, Kawamoto E (1995) Onshore tsunami deposits caused by the 1993 Southwest Hokkaido and 1983 Japan Sea Earthquakes. *Pure Appl Geophys* 144:693–717
- Switzer AD, Burston JM (2010) Competing mechanisms for boulder deposition on the southeast Australian coast. *Geomorphology* 114:42–54
- Tappin DR (2007) Sedimentary features of tsunami deposits – Their origin, recognition and discrimination: an introduction. *Sediment Geol* 200: 151–154
- Tuttle MP, Ruffman A, Anderson T, Hewitt J (2004) Distinguishing tsunami from storm deposits in eastern North America. The 1929 Grand Banks tsunami versus the 1991 Halloween storm. *Seismol Res Lett* 75:117–131
- Williams HFL, Hutchinson I (2000) Stratigraphic and microfossil evidence for late Holocene tsunamis at Swantown Marsh, Whidbey Island, Washington. *Quatern Res* 54:218–227

Tsunami from the Storegga Landslide

Stein Bondevik

Department of Environmental Sciences, Western Norway University of Applied Sciences, Sogndal, Norway

Article Outline

Glossary

Definition of the Subject

Introduction

How the Storegga Tsunami Was Discovered

The Slide That Triggered the Tsunami

Storegga Tsunami Deposits and Run-Up

Numerical Simulations of the Storegga Tsunami

Dating the Storegga Event

Storegga Tsunami and Stone Age Humans

Future Directions

Bibliography

Glossary

Mesolithic Refers to the middle period in the Stone Age. “Meso” means middle and “lithic” means stone or rock. The Mesolithic began with the Holocene warming, around 11,500 years Before Present (BP) and ended when farming was introduced in Scandinavia around 6000 years BP. The Storegga Slide happened in the Late Mesolithic when humans subsisted on fishing and hunting and the population had become more sedentary. Numerous excavations in Norway and Scotland have found Late Mesolithic settlements on or close to the former shores; these may have been affected by the Storegga tsunami.

Radiocarbon ages A radiocarbon year varies in length relative to a calendar year because the

amount of radioactive carbon, ^{14}C , in the atmosphere varies over time. Best age estimate of the Storegga tsunami is 7320 ± 20 radiocarbon years BP. By counting back tree rings year by year that have been systematically radiocarbon dated, this age would correspond to the tree rings that grew in the time interval 8120–8175 years BP. In the discussion to follow, radiocarbon years are denoted as ^{14}C year BP.

Retrogressive slide motion A retrogressive slide starts in the lower area of the slope and retreats backward, up the slope. A piece or block of deposits detaches and slides down-slope. The detached piece causes loss of support to the deposits behind it (upslope) causing another piece to detach. This development continues up the slope with the piecemeal release of other pieces, until the last piece on top of the slope has failed. A retrogressive slide motion is a typical development in submarine slides and in quick clay slides onshore where the slope angle is low. Such slides can often spread over long distances.

Run-up Maximum vertical height the tsunami runs up a slope. It is measured in meters from the sea level at the time the tsunami hits and up, vertically, to the maximum point of inundation. Reconstructions of run-up from ancient tsunamis are difficult for two reasons: first, you need to find the highest reaching tsunami deposit; and, second, you need to reconstruct the sea level when the tsunami hit. Run-up reconstructed from tsunami deposits is usually minimum values and could be called sediment run-up.

Sediment trap A place where sediments accumulate and form deposits. Tsunamis are mainly erosive and tend to wash away vegetation, soil, and loose deposits into the ocean. In order to find deposits left onshore, there has to be some kind of a trap in which sediments carried by the tsunami are deposited and later protected from other processes, like running water, wind, burrowing animals, and human activity.

Coastal lakes, estuaries, and bogs are such traps where Storegga tsunami deposits have been preserved.

Definition of the Subject

The Storegga tsunami was generated by the Storegga landslide off the Norwegian coast about 8150 years BP. The tsunami deposits show that the coasts of Scotland, Norway, Shetland, Faroe Islands, and possibly also Eastern Greenland and Denmark were inundated and that the tsunami ran up to heights ranging from 3 to more than 20 m above sea level of that time. The Storegga tsunami is important for two reasons: First, it shows that big tsunamis have happened along passive margins and outside of the Pacific Ocean region. Second, it is the only slide-generated tsunami of a basin-wide scale where the run-up has been well mapped in the field and the tsunami waves simulated with numerical models.

Introduction

When geologists first studied deposits from the Storegga tsunami, they interpreted them as resulting from other processes, such as a flood, a big storm surge, or a sea level rise. The Norwegian Sea and North Sea have passive margins with small earthquakes, so to think of tsunami deposits in this part of the world was not in anyone's mind until the Storegga Slide was discovered in the 1980s. Moreover, the tsunami-sediment layer was often overlooked when encountered in the field during sea level studies because it was just "noise" in the overall picture and in field notes was often mentioned as some kind of disturbance that complicated the local stratigraphy.

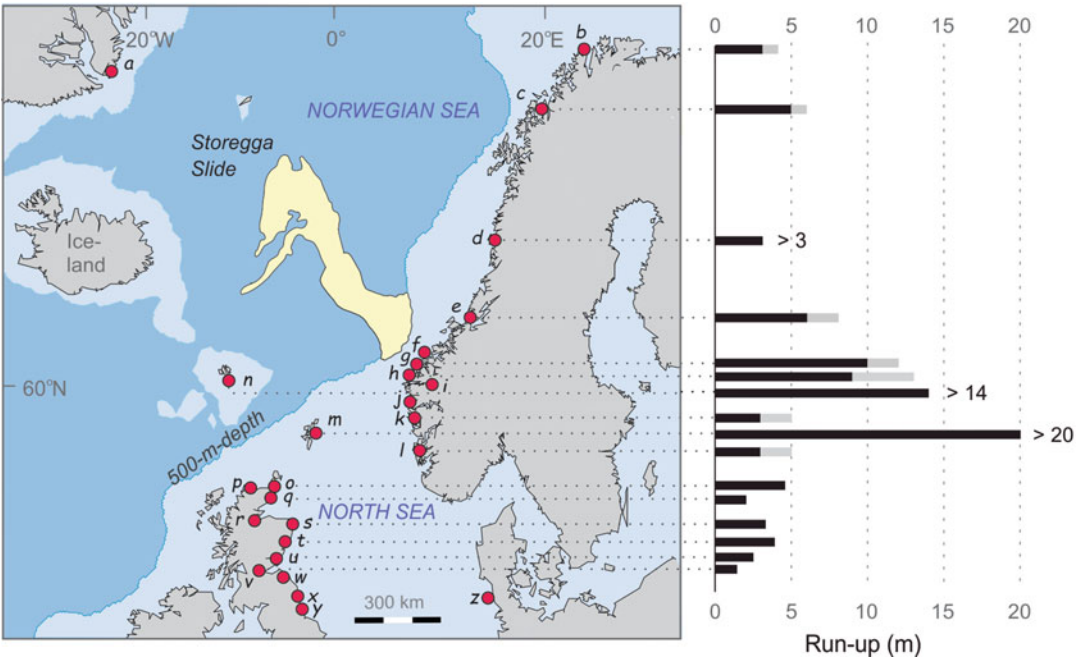
Run-up pattern (Fig. 1) and radiocarbon dates point toward the Storegga Slide as the source for the tsunami. Sites closest to the slide, like Western Norway, Shetland, and Faroe Islands, show the largest run-up – ranging from about 8 to more than 20 m (Fig. 1). Sites farther away show run-up less than 5 m (Scotland, Northern Norway). The most

accurate date for the tsunami is 8150 ± 30 years BP and comes from radiocarbon ages of moss, unexpectedly well preserved within the tsunami deposits (see section "[Dating the Storegga Event](#)") that were killed during the tsunami event. The slide itself was dated from radiocarbon ages of calcareous foraminifera found within the first centimeters of sediments deposited on top of the slide material. Those foraminifera ages fall within the time interval of 8100 ± 250 years BP. Both the run-up pattern and radiocarbon dates suggest that the tsunami was triggered by the Storegga Slide.

The Storegga tsunami deposits are preserved in various kinds of "sediment traps": coastal lake basins, estuaries, and bogs. These different sites are listed in Table 1 together with references to the original publications, type of sediment traps, run-up (if reconstructed), and short comments about the run-up estimates and/or deposits. In the following text, I will present and discuss the findings from one location with lake and marine basins (Western Norway), one location from an estuary (Scotland), and one from a peat outcrop (Shetland). For information about the other sites, the reader will need to consult the publications referenced in Table 1. First however, I will summarize how the Storegga tsunami was discovered and then describe the slide that generated the tsunami.

How the Storegga Tsunami Was Discovered

"Storegga" (meaning the "big edge") was the name originally given to about a 100-km-long stretch of a prominent undersea feature, with a total length of ~300 km, in the Norwegian Sea (Figs. 2 and 3). Along this edge, fishing can be very good because of strong upwelling, and men from Western Norway have been fishing here for at least 300 years (Rødal 1996). The anglers found a big change in depth – from the shallow shelf of about 200–300 m depth to 500–600 m depth within a very short distance; thus, they named the place *Storegga*. Little did they know that their fishing place was along the back wall to



Tsunami from the Storegga Landslide, Fig. 1 The Storegga Slide in the Norwegian Sea is among the world's largest submarine landslides. Red dots show locations of Storegga tsunami deposits; letters correspond to the same

sites described in Table 1. Run-up estimates, inferred from deposits, are given to the right; black columns show minimum estimates, and gray columns give maximum estimates

what has been called the largest submarine landslide in the world (Bugge et al. 1987; Hafliðason et al. 2005).

The slide itself was identified when the continental shelf of Norway was mapped in the 1970s (Bugge 1983; Bugge et al. 1978). It was soon recognized as a very big slide complex, thought to consist of three individual slide events (Bugge et al. 1987; Jansen et al. 1987). However, much more detailed data (e.g., multi-beam echo sounder, detailed high-resolution seismic lines, and sediment cores) showed that the slide was not three individual events as initially believed but instead was one major event consisting of different phases (Bryn et al. 2005a). The slide involved a volume of 2400–3200 km³ (Table 2), and from a large number of cores, the slide was radiocarbon dated to 8100 ± 250 years BP (Hafliðason et al. 2005). In the early paper by Jansen et al., (1987), it was speculated that such a big slide could have generated a tsunami.

Deposits of the Storegga tsunami were first recognized in Norway and Scotland in the 1980s. John Inge Svendsen, a student at that time at the University of Bergen, Norway, cored coastal lakes on Sunnmøre (site **h** in Fig. 1), and in one of the basins, he found a sand layer with brackish diatoms interbedded in freshwater lake mud. Alastair Dawson, at the University of Coventry, UK, knew of a widespread sand layer in estuarine mud in North East Scotland previously interpreted as a result of a storm surge (see below). Both suggested, independently, that these deposits were laid down by a tsunami generated from the Storegga Slide (Dawson et al. 1988; Long et al. 1989b; Svendsen 1985; Svendsen and Mangerud 1990). Their interpretation was motivated by the discovery of the Storegga Slide; Dawson read the paper by Jansen et al. (1987) in Marine Geology about the Storegga Slide, and Svendsen heard first about the Storegga Slide from colleagues and fellow students at the same department studying the Storegga Slide (Befring 1984; Edwin 1984).

Tsunami from the Storegga Landslide, Table 1 Field observations of Storegga tsunami deposits and run-up estimates

Index on map	Area	Location	No of sites	Reference(s)	Outcrop	Lake basin	Marine basin	Estuarine mud	Peat	Run-up (m)	Comments
a	East Greenland	Geographical society Ø (Loon lake)	1	Wagner et al. (2007)			x			–	Sandy sequence with erosive base in marine silt interpreted as Storegga tsunami deposits. Dated to younger than 8500–8300 year BP
b	Northern Norway	Finnmark (Sorøya, Rolvsøya, Nordkinn)	5	Romundset and Bondevik (2011)		x				3–5	Coastal lakes along the outer coast of Finnmark show Storegga tsunami to propagate into the Barents Sea. Deposits in lakes 0–3 m above contemporary sea level. No traces in lake 5 m above contemporary sea level
c	Northern Norway	Troms (Lyngen)	3	Rasmussen et al. (2018)		x	x			6.5–7	Storegga tsunami barely inundated lake at 25.9 m a.s.l. Mean sea level at 19–19.5 m a.s.l
d	Northern Norway	Brønnøy (Hommelstø)	3	Bondevik et al. (2012) Rydgren and Bondevik (2015) Bondevik (2003a)			x			>3	All basins investigated were below sea level at Storegga time, terrestrial plants, and peat clasts within tsunami deposits indicate run-up >3 m
e	Mid-Norway	Bjugn	4	Bondevik et al. (1997a)		x	x			6–8	Storegga tsunami overflowed lake at 42 m a.s.l., but not lake at 44 m a.s.l. Contemporary high tide sea level at ca. 36 m a.s.l
f	Western Norway	Sunnmøre (Harøy)	1	Bondevik (2003b)	x				x	–	Sand layer in peat below Tapes beach ridge was traced in cores landward of the beach ridge
g	Western Norway	Sunnmøre (Sula)	6	Bondevik et al. (1997a)		x	x			10–12	Storegga tsunami barely inundated lake at 21.5 m a.s.l., but no traces of inundation in lake at 22 m a.s.l. Contemporary sea level at 10–11 m a.s.l
h	Western Norway	Sunnmøre (Bergsøy, Leinøy)	4	Bondevik et al. (1997a) Svendsen (1985) Svendsen and Mangerud (1990)		x	x			9–13	A lake 8–9 m above contemporary sea level shows large erosion from Storegga tsunami. Two lakes 12–13 m above contemporary sea level have no traces of inundation
i	Western Norway	Nordfjord	2	Vasskog et al. (2013)		x				>1–5	Two lakes at the head of the fjord “Nordfjord” – up to 3 m thick Storegga tsunami deposits. Lakes were 1–5 m above contemporary high tide sea level
j	Western Norway	Florø	1	Aksdal (1986) Bondevik et al. (1997a)		x				–	A bed of gravel, sand, and redeposited gyttja, draped with 2 mm silt, dated to 7360 ± 110 ¹⁴ C years
k	Western Norway	Hordaland (Austrheim)	5	Bondevik et al. (1997a)		x				3–5	A lake at 14 m a.s.l. was clearly inundated by the Storegga tsunami, but not

(continued)

Tsunami from the Storegga Landslide, Table 1 (continued)

Index on map	Area	Location	No of sites	Reference(s)	Outcrop	Lake basin	Marine basin	Estuarine mud	Peat	Run-up (m)	Comments
											a lake at 15 m a.s.l. Contemporary sea level at 10–11 m a.s.l.
l	Western Norway	Hordaland (Bømlo)	2	Bondevik et al. (1997a)		x				3–5	Storegga tsunami deposit in lake at 15 m a.s.l, but not in a bog at ca. 16 m a.s.l. Contemporary sea level at ca 12 m a.s.l
m	Shetland	Unst, Sullom Voe	9	Bondevik et al. (2005b) Smith et al. (2004)	x	x			x	>20	Storegga tsunami deposit found in 4 lakes located 0.5–3 m above present high tide level. Sand layer in peat outcrops was traced to 9.2 m above high tide on the western side of the Sullom Voe area and to 11.8 m above spring tide on the eastern side. Contemporary sea level 10–15 m below present sea level
n	Faeroe Islands	Suderøy (Vágur)	1	Grauert et al. (2001)		x				>14	Lake at 4 m a.s.l. with Storegga tsunami deposits, at the head of 5-km-long fjord. Contemporary sea level probably 10 m below present sea level
o	Scotland	Sutherland, Caithness	2	Dawson and Smith (1997, 2000) site 7 & 8 in Smith et al. (2004)		x		x	x	4.6	In lower Wick River valley, Caithness, fine sand layer within peat. Tsunami deposits in filled-in lagoon at Strath Halladale in Sutherland
p	Scotland	Loch Eriboll, Lochan Harvurn	1	Long et al. (2016)	x				x	–	Sand layer in a coastal cliff section of peat
q	Scotland	Dornoch Firth	3	Site 9–11 in Smith et al. (2004) and references therein				x		2	Widespread sand layer (Figs. 13 and 14)
r	Scotland	Inner Moray Firth	6	Smith et al. (2004) and references therein. Dawson et al. (1990)				x	x	3.3	Layer of fine sand of marine provenance in estuarine sediments; landward it rises into peat. One of the sites is an archeological excavation at Inverness, “beach sand” resting upon a Mesolithic horizon with artifacts
s	Scotland	North East Scotland	3	Smith et al. (2004) and references therein				x	x	3.3	Fine-medium sand within peat landward, seaward it continuous into estuarine mud
t	Scotland	Tayside	6	Smith et al. (2004) and references therein. Dawson et al. (1988)	x			x	x	3.9	Fine sand layer in estuarine deposits around the Montrose Basin, continuous into peat landward. One site (Maryton) is in a cliff exposure (Fig. 15)
u	Scotland	Near St. Andrews (Silver Moss, Craigie)	2	Smith et al. (2004) Dawson et al. (1988)				x	x	2.9	Tapering layer of sand in estuarine deposits that pass into peat upslope

(continued)

Tsunami from the Storegga Landslide, Table 1 (continued)

Index on map	Area	Location	No of sites	Reference(s)	Outcrop	Lake basin	Marine basin	Estuarine mud	Peat	Run-up (m)	Comments
v	Scotland	Firth of Forth	2	Smith et al. (2004) and references therein				x	x	1.4	Sand within estuarine mud passes into peat at the valley side
w	Scotland	Near Dunbar (East Lothian)	1	Smith et al. (2004) and references therein					x	1.3	Sand in peat moss that contains marine and brackish diatoms. Radiocarbon dated to between 7590 ± 60 and 7315 ± 70 ^{14}C year BP
x	NE England	Broomhouse Farm	1	Shennan et al. (2000) Smith et al. (2004)					x	ca. 3	Sand horizon of marine provenance within coastal peat moss at Broomhouse Farm
y	NE England	Howick, Northumberland	1	Boomer et al. (2007)				x		–	30 cm layer of coarse sands and pebbles dated to 8300 year BP in marine clay/silt
z	Denmark	Rømø	1	Fruergaard et al. (2015)		x				–	80 cm thick layer of sand and rip-up clasts of organic material

Many researchers had encountered the Storegga tsunami deposits earlier, but, without knowing about the Storegga Slide, they explained the deposits as something else. The first scientist to core lake basins in Western Norway, Knut Fægri, probably saw traces of the Storegga tsunami but interpreted them as representing a sea level fluctuation (Fægri 1944). Sissons and Smith (1965) were the first to study the layer in Scotland, where they found a persistent 5-cm thick sand layer in estuarine mud and peat, which they thought was deposited from a local flood of a river (Smith et al. 2004). Later, after discovering a similar layer at other sites in eastern Scotland, Smith et al. (1985) suggested it was caused by a “storm surge of unusual magnitude.” In Norway, the coarse-grained layer found in lake mud was often interpreted to represent the peak of the mid-Holocene transgression (called the Tapes transgression in Norway) that happened about the same time (Aksdal 1986; Corner and Haugane 1993; Kaland 1984; Svendsen and Mangerud 1987; Tjemsland 1983) and that the layer was the result of strong currents running through the basins as they became submerged during this marine transgression (Kaland 1984).

The deposit was soon mapped and studied along the eastern coast of Scotland (Dawson et al. 1988;

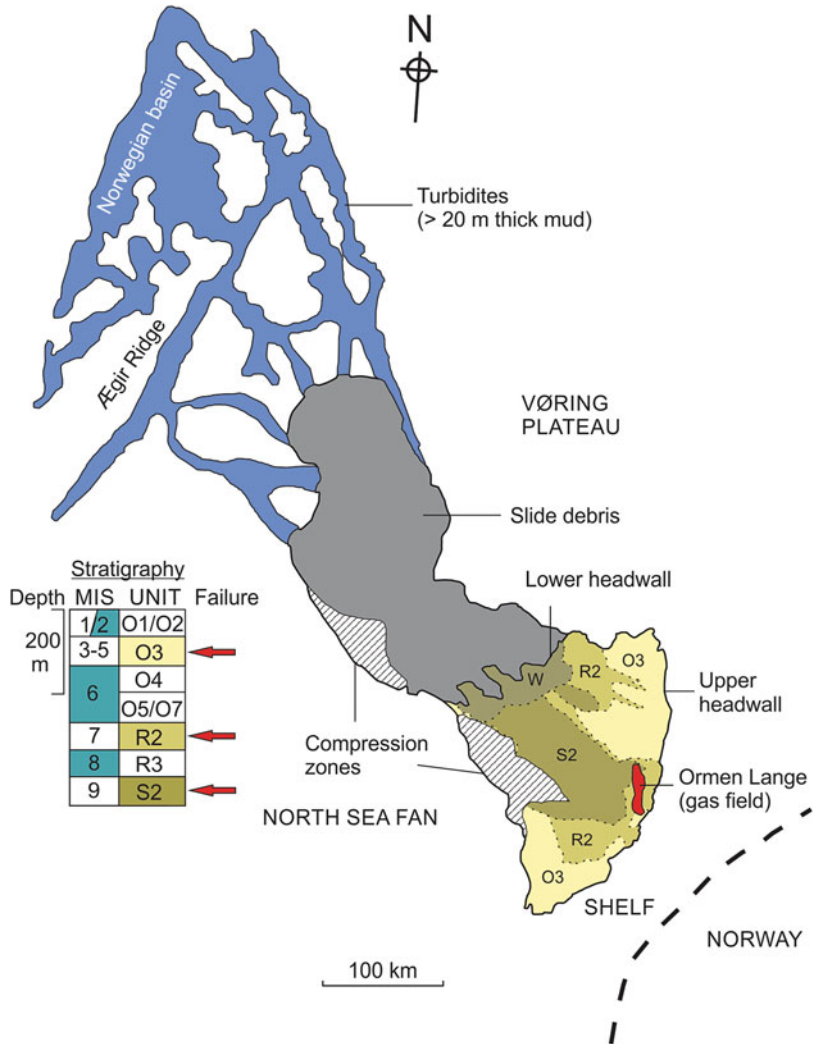
Long et al. 1989b; Smith et al. 2004), in Western Norway (Bondevik et al. 1997a,b), the Faeroe Islands (Grauert et al. 2001), and Shetland (Bondevik et al. 2003, 2005b; Smith et al. 2004). During these early studies, the authors tried to answer the question whether this deposit could be the Storegga tsunami deposit and focused on the layer’s age and stratigraphical context (Dawson et al. 1988; Long et al. 1989b; Svendsen and Mangerud 1990). Later studies focused on the sedimentology of the deposits (Bondevik et al. 1997b; Dawson et al. 1991; Dawson and Smith 2000; Shi 1995), run-up (Bondevik et al. 1997a; Smith et al. 2004), and numerical modeling of the tsunami waves (Bondevik et al. 2005a; Harbitz 1992; Hill et al. 2014; Løvholt et al. 2017).

The Slide That Triggered the Tsunami

The Storegga Slide is huge. It covers an area of 95,000 km² – the size of a European country like Portugal (92,391 km²) – and involved a total volume of 2400–3200 km³ (Haflidason et al. 2005). You could almost fit Belgium (30,528 km²) in the slide scar (Table 2, Figs. 2 and 3a). The volume is so big it is hard to imagine, but if you distribute the volume across the USA

Tsunami from the Storegga Landslide,

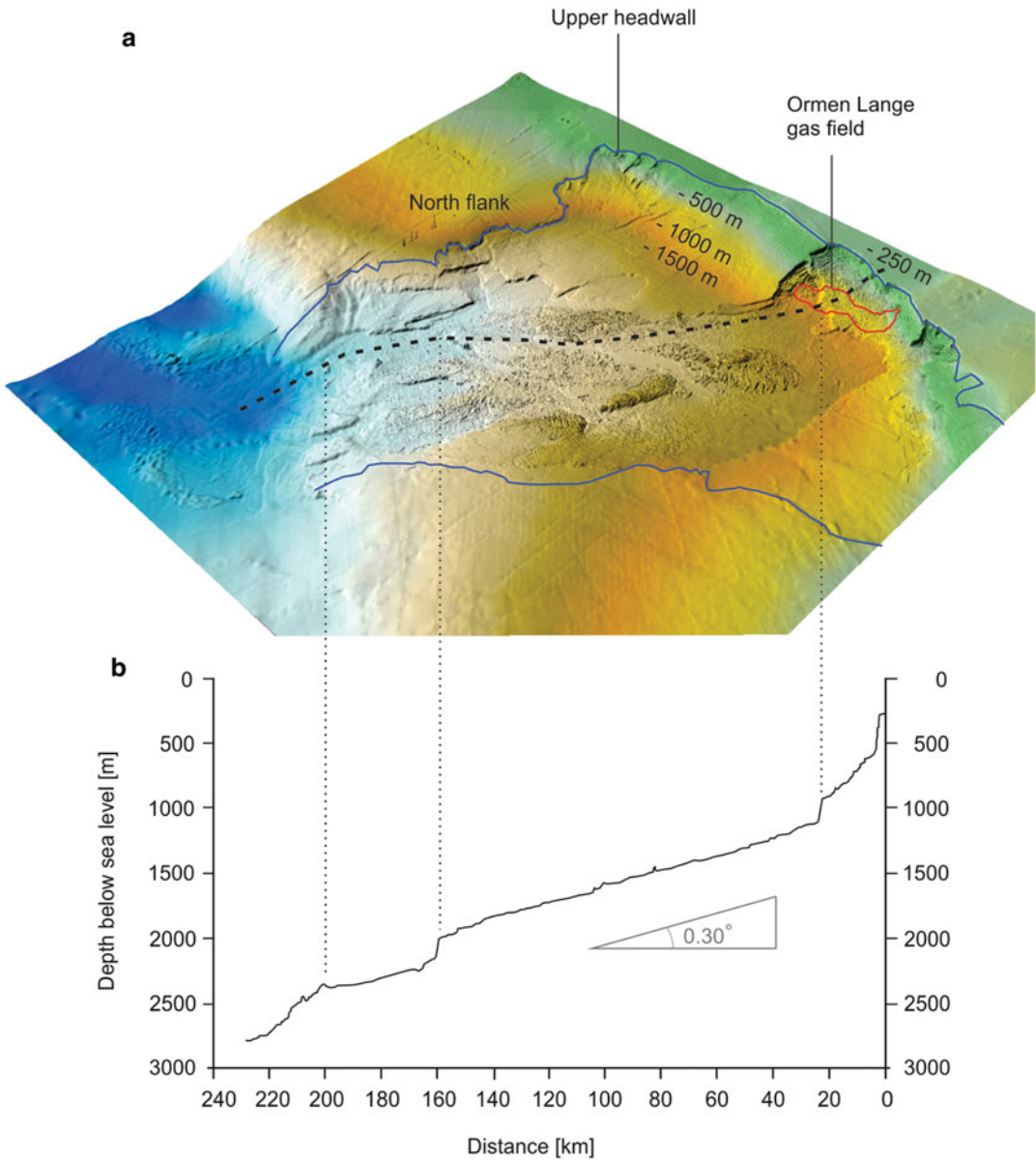
Fig. 2 Outline and deposits of the Storegga Slide. Failure layers in the slide scar area, between the lower and upper headwall, are contourites deposited during interglacials and interstadials (S2, R2, and O3). Dotted lines show minor headwalls. The stratigraphy in the area correlates to the Naust formation (last three My) and is subdivided into five units: W, U, S, R, and O (youngest) (Berg et al. 2005); unit U is not exposed in the Storegga scar area. The table to the left shows how the stratigraphic units relate to the marine isotope stages (MIS). (Based on Fig. 1 in Hafliðason et al. 2005 and Fig. 8 in Bryn et al. 2005a)



(9,834,000 km²), it will cover the land with a 30 cm thick layer! The back wall has a length of 310 km, and the total runout length – including the distal turbidites – is 810 km (Table 2, Fig. 2). Much of the slope of the slide is only between 0.3° and 2° (Fig. 3b), and a big question is how can anything fail on such a low angle slope and subsequently develop into a gigantic slide?

We learned much about the Storegga Slide in the Ormen Lange project – a project for the safe development of a deepwater gas field within the Storegga Slide complex (Solheim et al. 2005b). The “Ormen Lange” gas field – named after a Viking ship from its look-alike in map view

(Figs. 2 and 3a) – is located just beneath the major headwall 1900 m below the seafloor at a water depth of between 600 and 1200 m (Bryn et al. 2007). Major questions for the project to solve were whether a new tsunami-generating slide could occur in the area and if the planned activities and field installations to develop the gas field could trigger a new slide. To solve these questions, abundant data were gathered in the 1990s and the early 2000s, including high-resolution bathymetric and seismic data, drillings, numerical models of the slide dynamics, and trigger mechanisms. The short answer to these questions from the Ormen Lange project was that there



Tsunami from the Storegga Landslide, Fig. 3 (a) Bathymetric image of the slide scar. The Ormen Lange gas field is located close to the upper headwall (Redrawn from Fig. 2 in Kvalstad et al. 2005). (b) Depth profile along the dotted line in (a). The line of the slope surface is

“wavy” because of the large blocks in the slide. Note the minor headwalls along the profile – this is a jump to a glide plane at a higher stratigraphic level. A large part of the slope is only 0.3°. (Redrawn from Fig. 3 in Kvalstad et al. 2005)

is a very low risk for a new slide to happen today or in the near future at Storegga (Bryn et al. 2005a). The gas field commenced production in 2007.

The materials that failed in the Storegga Slide were glacial deposits and contourites of Quaternary age (Fig. 2). During periods of peak glaciation, the Scandinavian ice sheet would reach the

Tsunami from the Storegga Landslide, Table 2 Storegga Slide information

Total volume	2400–3200 km ³
Total area, including depositional area	95,000 km ²
Area of slide scar	27,000 km ²
Length of upper headwall	310 km
Water depth at upper headwall	150–400 m
Runout length, including distal turbidites	810 km
Water depth at distal-most deposits	3800 m
Average slope gradient	0.25°
Steepest gradient in upper headwall	35°
Max vertical height of upper headwall	250 m
Age of foraminifera on top of slide deposits	7250 ± 250 ¹⁴ C year BP; 8150 ± 250 years BP

Data from Hafliðason et al. (2005) and Solheim et al. (2005a,b)

shelf break and deposit till on the shelf and debris flows on the upper continental slope. This material deposited from the glacier consists of about equal amounts of clay, silt, and sand and has low water content (10–20%) and high density. When the ice front was in a retreating position, during deglaciation and in interglacials and interstadials, silty clay with high-water content (25–35%) and lower density would be deposited (Berg et al. 2005). This silty clay was/is deposited underneath the North Atlantic Current that runs along the continental slope and is called “contourites,” a name derived from “contour” lines. Thus, the stratigraphy in the upper part of the Storegga area is a result of the climate cycles in the Quaternary consisting of alternating layers of glacial deposits and hemipelagic silty clay – the contourites (Bryn et al. 2005b).

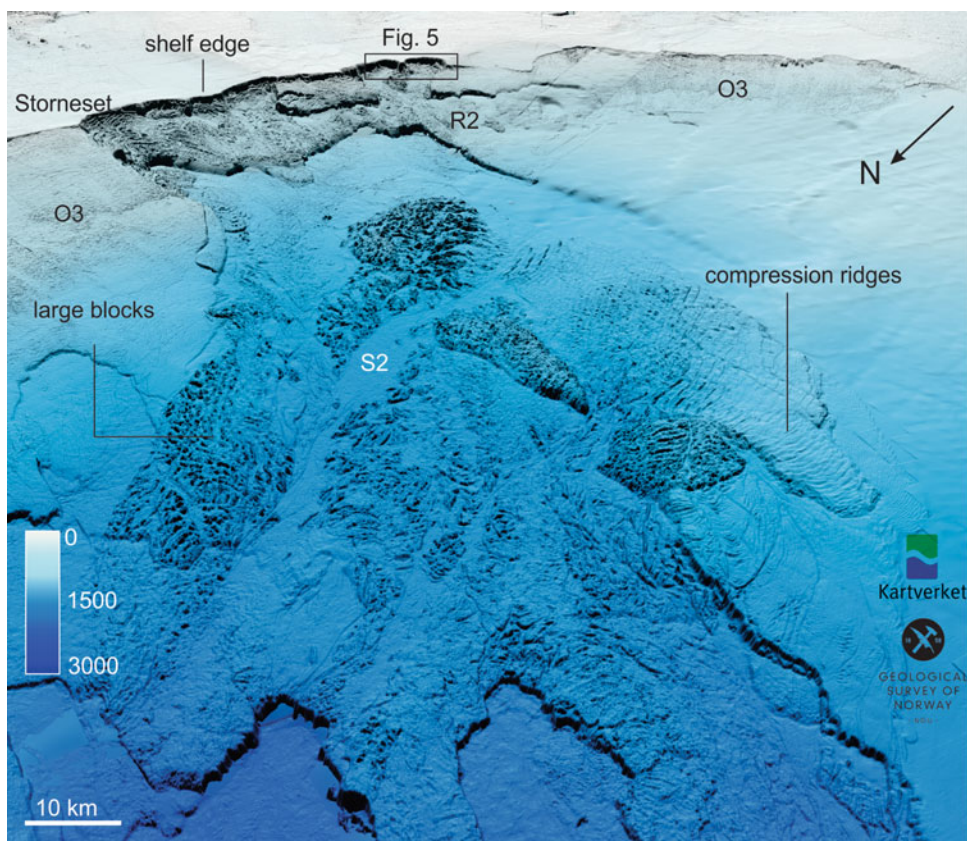
The slip planes or failure zones are all found within the contourites, the fine-grained silty clay. These sediments have different geotechnical behavior than the glacial deposits. Upon triaxial testing they develop clear shear planes and reach peak strength, and when further deformed their strength is substantially reduced, a process called

strain softening or brittle deformation (Berg et al. 2005). The scarps or minor headwalls in the Storegga Slide (Figs. 2 and 3) indicate a glide plane jump to a higher stratigraphic level of a younger contourite. Within the northern and southern parts of the slide scar, the slide is less than 100 m thick and slid along the O3 unit – a contourite formed during isotope stages 3–5. In the central part of the slide scar, the slip has followed deeper contourite layers, involving the R2 (isotope stage 7) and S2 (isotope stage 9) layers (Figs. 2, 3, 4, and 5).

It is difficult to explain how the Storegga Slide could happen because of the very low slope gradient (Fig. 3b). Seismic profiles and detailed bathymetry show it must have developed retrogressively (Hafliðason et al. 2004). This means that the slide started somewhere in the lower part of the slope and retreated upslope. Big pieces of slope material were released one at a time, and the headwall would gradually retreat upslope during the slide process. This kind of mass movement, where the overlying unit breaks into blocks that slide on a gentle slip surface, is called lateral spreading (Micallef et al. 2007). Because of the low slope gradient, the blocks must have slid on a very weak and soft layer without much shear strength (Kvalstad et al. 2005).

The clay layers that developed into failure planes lost their strength through strain softening. The unloading of the toe or lower headwall caused expansion of the slope material, and large strain developed in the softer clay, contourites, below the denser glacial material. The strain caused so-called strain softening of the contourite layer – a process where the strength of the material drops as the stress increases. Modeling has shown that such a process is possible where unloading of the headwall causes strain concentrations and strength loss in the failure layer sufficient to reduce the factor of safety below 1 and thus initiates a retrogressive slide process (Kvalstad et al. 2005).

The retrogressive failure mechanism requires the triggering of the first slide in the toe area. Factors that would lead to an increase in pore pressure and thus reduce the shear strength have been considered, and the most probable triggers are



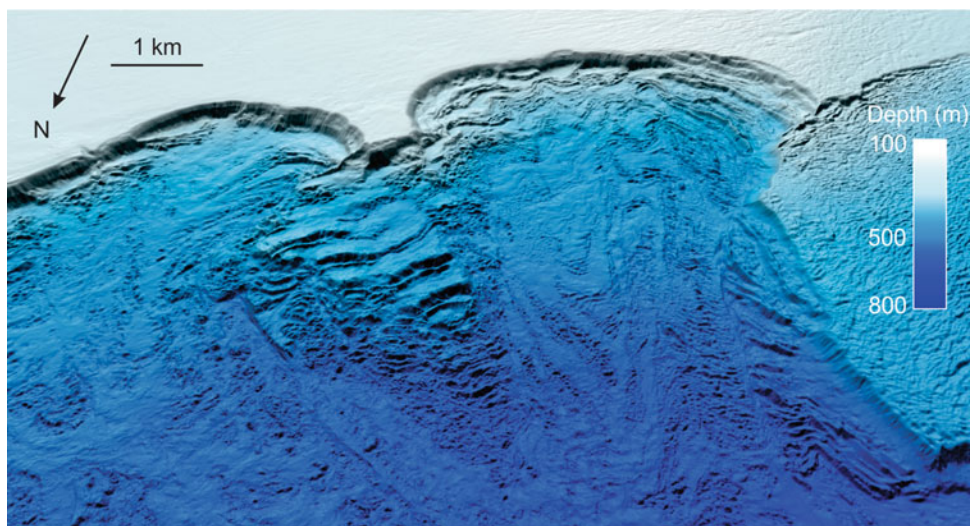
Tsunami from the Storegga Landslide, Fig. 4 Bathymetric image of the central part of the slide scar, where the slide cuts deepest into the deposits. Units

O3, R2, and S2 acted as glide planes. Terrain data: MAREANO/NHS, 3D visualization: MAREANO/NGU (www.mareano.no)

high sedimentation during peak glaciation, gas hydrate melting, and/or an earthquake (Bryn et al. 2005a). Rapid loading of glacial debris with low permeability may cause development of excess pore pressure. However, the sedimentation rate at the lower part of the Storegga slope is low. One mechanism that could lead to instability would be migrating pore water from the North Sea fan into the toe area – estimated to be about 160 km down-slope of the major headwall (Fig. 2). The North Sea fan was a major depocenter during the glacial maximum, and one idea is that pore water from this fan could be pushed into the lower part of the Storegga area and increased the pore pressure in this part of the slope and in that way reduced the shear strength (Bryn et al. 2005a).

Storegga Tsunami Deposits and Run-Up

Tsunami deposits are only rarely preserved, and traces of ancient tsunamis can be hard to find. A prerequisite is a sediment trap where sediments have been left by the tsunami and are protected from, e.g., wind, burrowing animals, running water, and human activity. The traps where Storegga tsunami deposits have been found are lakes and shallow marine basins near the coast, estuaries, or bogs – low areas where peat is accumulated. The differences between the dark organic lake gyttja, estuarine mud or dark peat, and the light-colored tsunami sand have helped to distinguish and to map the tsunami deposits in the field.



Tsunami from the Storegga Landslide, Fig. 5 Detailed image of the two “amphitheatres” along the upper headwall (for location see Fig. 4). Terrain data: MAREANO/NHS, 3D visualization: MAREANO/NGU (www.mareano.no)

In order to reconstruct tsunami run-up heights, the sea level at Storegga time had to be determined for each site (Table 1). For the coasts of Norway, Scotland, and Greenland, the shoreline 8150 years ago is located *above* present-day sea level. How much the 8150-year ago shoreline is above the present-day shoreline depends on the rate of isostatic uplift the site has experienced since deglaciation. The opposite is the case for the Faeroe and Shetland Islands, where the sea level has risen more than land since the deglaciation, and sea level at Storegga time was about 10–15 m *below* the shoreline of today. The accuracy of the run-up estimates depends on tracing the tsunami deposits to its highest elevation and determining the sea level at Storegga time correctly. It is also important to remember that the upslope limit of a tsunami deposit usually would be lower than the actual limit of the water in the tsunami (e.g., Smith et al. 2007). The true run-up could be several meters higher than the height of the tsunami deposits.

Coastal Lakes in Norway

The tsunami traps in Norway, Greenland, and the Faeroe Islands (a few in Shetland) are lake basins

close to sea level (Figs. 6 and 7; Table 1). These basins, most of them in bedrock, were excavated by glacial ice during the ice ages and have accumulated sediments since the last deglaciation. Because they are overdeepened, the lake floor deposits are usually protected from later erosion. The outer coast of Norway has many such basins (e.g., Fig. 9), and if located below the marine limit, they would hold a sedimentary record of sea level changes. Accurate sea level curves have been reconstructed from the deposits in such basins in Scandinavia (e.g., Svendsen and Mangerud 1987). Through such sea level studies, geologists in Norway encountered the Storegga tsunami deposits in lake basins at the coast, and in the beginning, they did not understand what they were (see section “How The Storegga Tsunami Was Discovered”).

The tsunami deposits form a distinct group of facies (Fig. 8) that are very different from the other sediments in the lakes (Figs. 8, 10, and 11). The tsunami deposits rest on an erosional unconformity that usually could be traced throughout the basins. Typically, we found more erosion at the end of the basin that faces toward the sea (Fig. 12b). The deposits can be divided into six different facies



Tsunami from the Storegga Landslide, Fig. 6 Coastal lake in Shetland, the Loch of Snarra Voe, on the Island Unst (Fig. 16). The lake is 0.6 m above the present high tide level but was probably at least 10 m above sea level when Storegga happened. Here we found distinct Storegga

tsunami deposits with gravel, sand, rounded rip-up clasts of well-consolidated silt, and different types of marine shell fragments. Most erosion was observed in cores between the raft and the outlet of the lake (Bondevik et al. 2005b)

(see Table 3 and Fig. 8). In some cores, we found all the six different facies (Fig. 11 has five of them), but at other places, only one or two of them is/are represented (Fig. 12c).

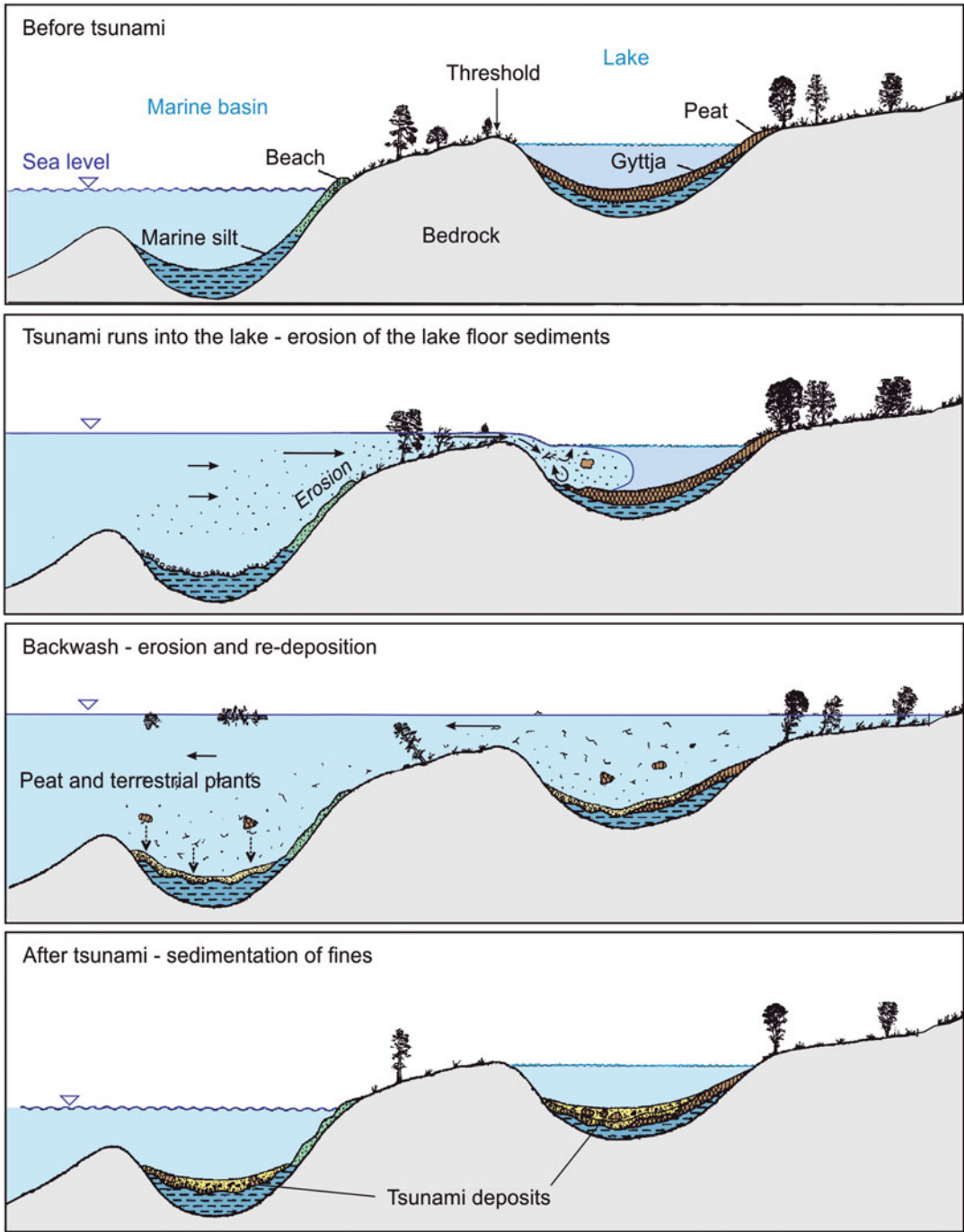
The main arguments that these facies were deposited by a tsunami were listed in Bondevik et al. (1997b) and are repeated here:

1. The geometry of the bed with seaward erosion (Fig. 12b) and landward fining (Fig. 12a) and the content of marine fossils clearly indicate a marine-related process for its formation.
2. Radiocarbon dates show the same age in all areas, independent of the elevation of the basin above sea level (Figs. 11 and 12d).
3. These facies are also found in basins that were below sea level at Storegga time

(Figs. 10 and 23), indicating that the backwash brought terrestrial material (rip-up clasts and terrestrial vegetation) into the sea.

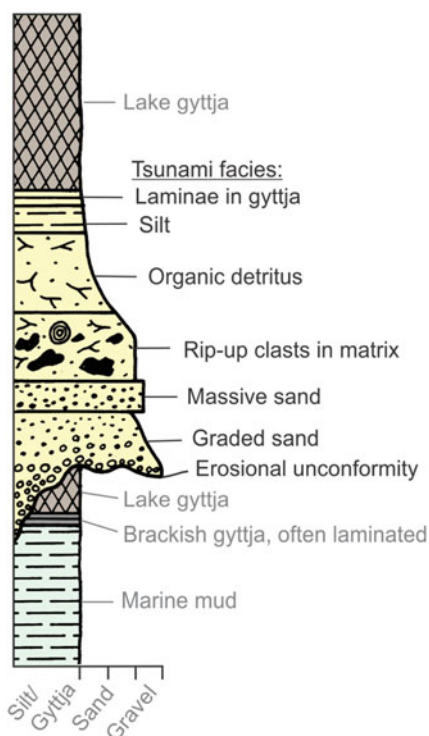
4. Many of the characteristics of these deposits are also reported from known modern tsunami deposits: extensive erosion, rip-up clasts, decrease in thickness and grain size landward, alternation between finer- and coarser-grained beds, and, as a whole, the deposit generally fines upward.

We think that the different sand layers in the tsunami deposits may represent successive incoming tsunami waves. In the lower basins, less than 3 m above the sea level at Storegga time, there may be several sand beds interbedded with organic detritus (Figs. 10 and 11), whereas in the



Tsunami from the Storegga Landslide, Fig. 7 A sedimentological model for erosion and deposition by a tsunami inundating a marine basin and a lake basin. The

model is based on analysis of Storegga tsunami deposits in Western Norway. (Redrawn from Fig. 12 in Bondevik et al. 1997b)



Tsunami from the Storegga Landslide. **Fig. 8** Description of the Storegga tsunami deposits (in yellow) presented as an idealized, complete facies sequence of tsunami deposits in nearshore lakes. The enclosing sediments are also shown. Note that if graded sand and massive sand occur in the same tsunami sequence, they are normally separated by organic deposits (rip-up clasts, detritus). (Based on Fig. 5 in Bondevik et al. 1997a)

higher basins, 5–10 m above contemporary sea level, only one sand bed is usually found (Fig. 12c). The individual sand layers may be deposited by the incoming waves and the organic muddy material in between (Fig. 11) settled out during slack water, the period between the waves. However, we cannot rule out that material deposited in the first inundating waves could have been later eroded away by the subsequent tsunami waves inundating the basins.

To determine the run-up at a location, we (e.g., Bondevik et al. 1997a; Romundset and Bondevik 2011) cored lake basins at different elevations and traced the tsunami deposits to the highest lake. The precision of these measurements depends on how closely the lakes are located relative to the true run-up heights and how accurate the sea level

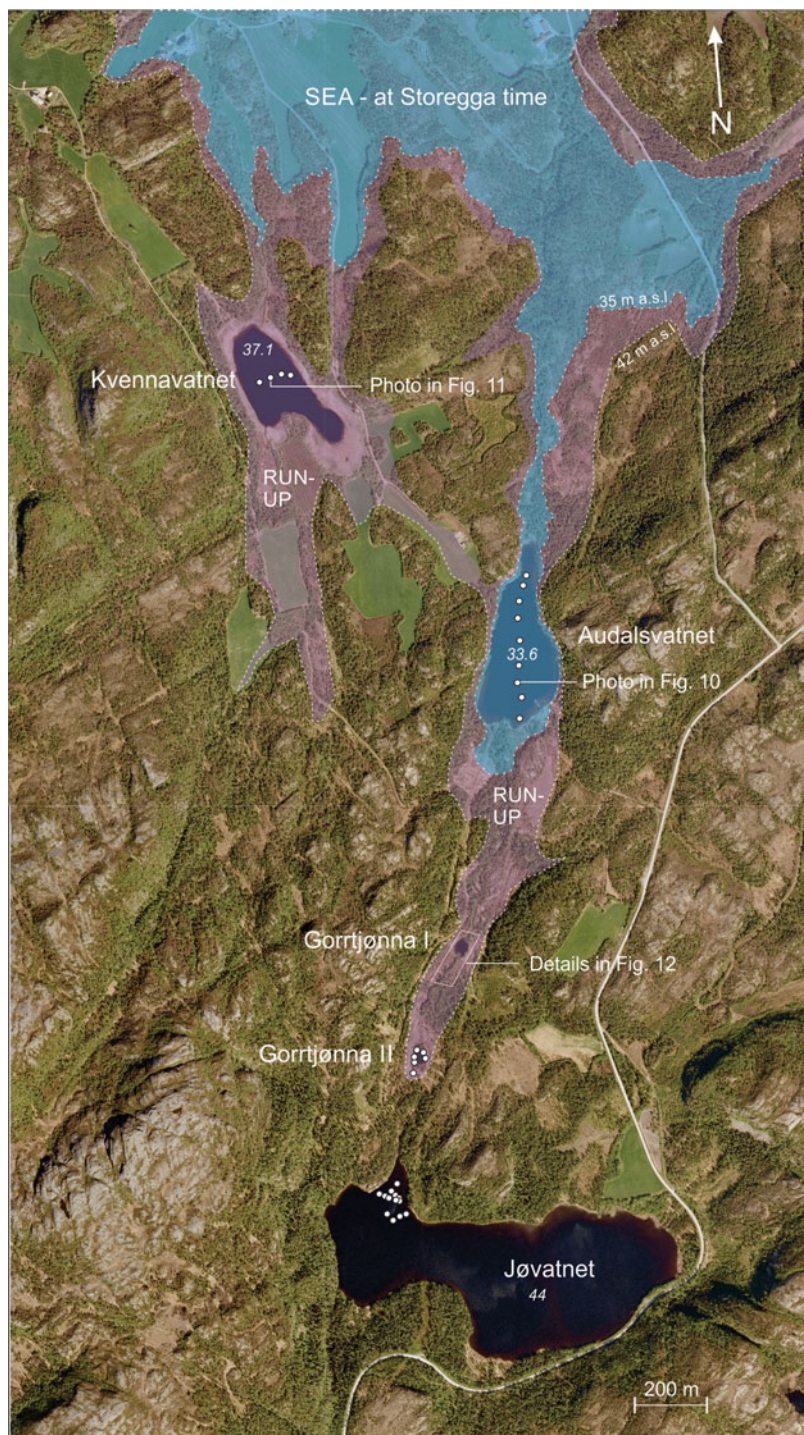
at Storegga time had been reconstructed for that area. If a tsunami deposit is found in a lake basin, it means that the tsunami had to flow over the outlet threshold of the lake (Fig. 7). A minimum estimate of the height of the tsunami would then be the altitude of the threshold of the basin relative to the local sea level at Storegga time. The tsunami deposits were then traced successively to lakes at higher elevations. The first lake without any traceable deposit that could indicate inundation was interpreted to not have been reached by the tsunami.

For example, at site e (Fig. 1, Table 1), in Bjugn, the run-up was reconstructed based on the deposits and thresholds of five lake basins (Fig. 9) (Bondevik et al. 1997a). Sea level was above lake Audalsvatnet at 33.6 m a.s.l. when the Storegga tsunami happened because the tsunami deposits are found within marine sediments (Figs. 9 and 10). In the deposits in Kvennavatnet (37.1 m a.s.l.), diatoms show a change from marine to brackish sediments just underneath the tsunami deposits, but in the gyttja above the tsunami deposits, the diatoms are solely freshwater species (Fig. 9 in Bondevik et al. 1997a). A moss stem from the brackish gyttja just below the tsunami deposits was dated to 7350 ± 80 ^{14}C years (Fig. 11) – close to the time of the Storegga event. This indicates that Kvennavatnet was very close to the sea level and had probably just emerged from the sea at Storegga time. We estimated that 36 m a.s.l. was a reasonable estimate for the high tide sea level at Bjugn at Storegga time.

The next two lakes higher up are Gorrjtjønn I and Gorrjtjønn II, both with tsunami deposits and with a threshold at 42 m a.s.l. These two small lakes are today separated by an area of peat growth (Fig. 9), but at Storegga time, they likely comprised one open lake basin. The stratigraphy in Gorrjtjønn I shows an episode of deep erosion, most erosion toward the threshold, and deposition of many rip-up clasts and gravel and sand that is finer grained toward land, away from the sea (Fig. 12). In Gorrjtjønn II, we found 1–2 cm layer of fine sand to silt and in few of the cores also with plant fragments; I now interpret this to be the distal deposits of the Storegga tsunami. A detailed coring program (with 13 cores, Fig. 9) at the outlet area of Jøvatnet (44 m a.s.l.) – where

Tsunami from the Storegga Landslide,

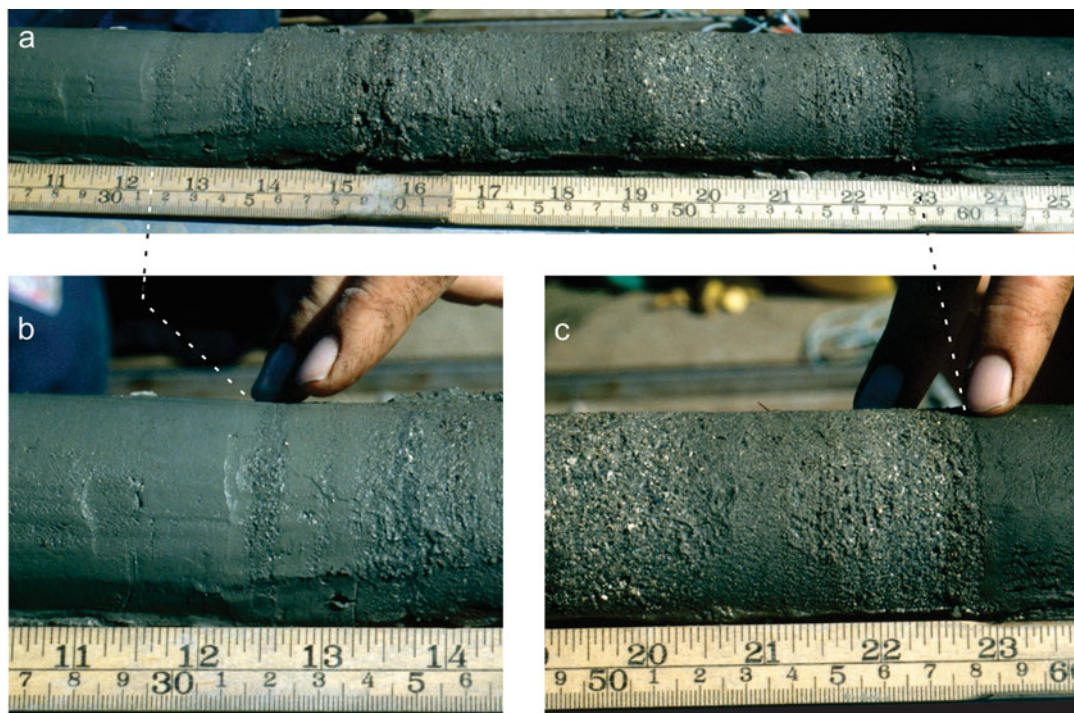
Fig. 9 Air photo of site e (Fig. 1, Table 1), Bjugn. Mean sea level at tsunami time in transparent blue color, here shown as the surface below the 35 m contour line. Run-up is in transparent violet color, the surface between the 35 m contour line and up to ca. 42 m contour line. Cores as white dots. No tsunami deposits were found in Jøvatnet (44 m a.s.l.). Audalsvatnet (33.6 m a.s.l.) was a few meters below sea level when the tsunami happened



we should expect the inflow of the tsunami – did not reveal any candidate for a tsunami layer.

At this site, a run-up of sediments is more than 6 m and less than 8 m. The Storegga tsunami

clearly inundated Gorrjønna I at 42 m a.s.l but not Jøvatnet at 44 m a.s.l. The local high tide sea level was estimated to 36 m a.s.l. Thus a minimum estimate is 42 m a.s.l. minus 36 m a.s.l. = 6 m, and



Tsunami from the Storegga Landslide, Fig. 10 (a) Photo of the Storegga deposits in Audalsvatnet, Bjugn (see location of core in Fig. 9). Ruler shows inches and cm. Below is a close-up of the upper boundary (b) and the lower boundary (c). The white spots in the sand are shell fragments. The four lowermost layers are graded sand; the

other two sand layers above are massive sand separated by silt. Terrestrial moss stems were found in the sand layers – one was dated to 7315 ± 70 ^{14}C years BP (Bondevik et al. 1997a). The lower boundary is knife-sharp; the upper boundary is gradual, somewhere around 28 cm (1228 cm below the lake surface) in gray silt

a maximum estimate is 44 m a.s.l. minus 36 m a.s.l. = 8 m. The inundation limit, from the 35 m contour line at Audalsvatnet to Gorrjtjøna II, is 900 m (Fig. 9).

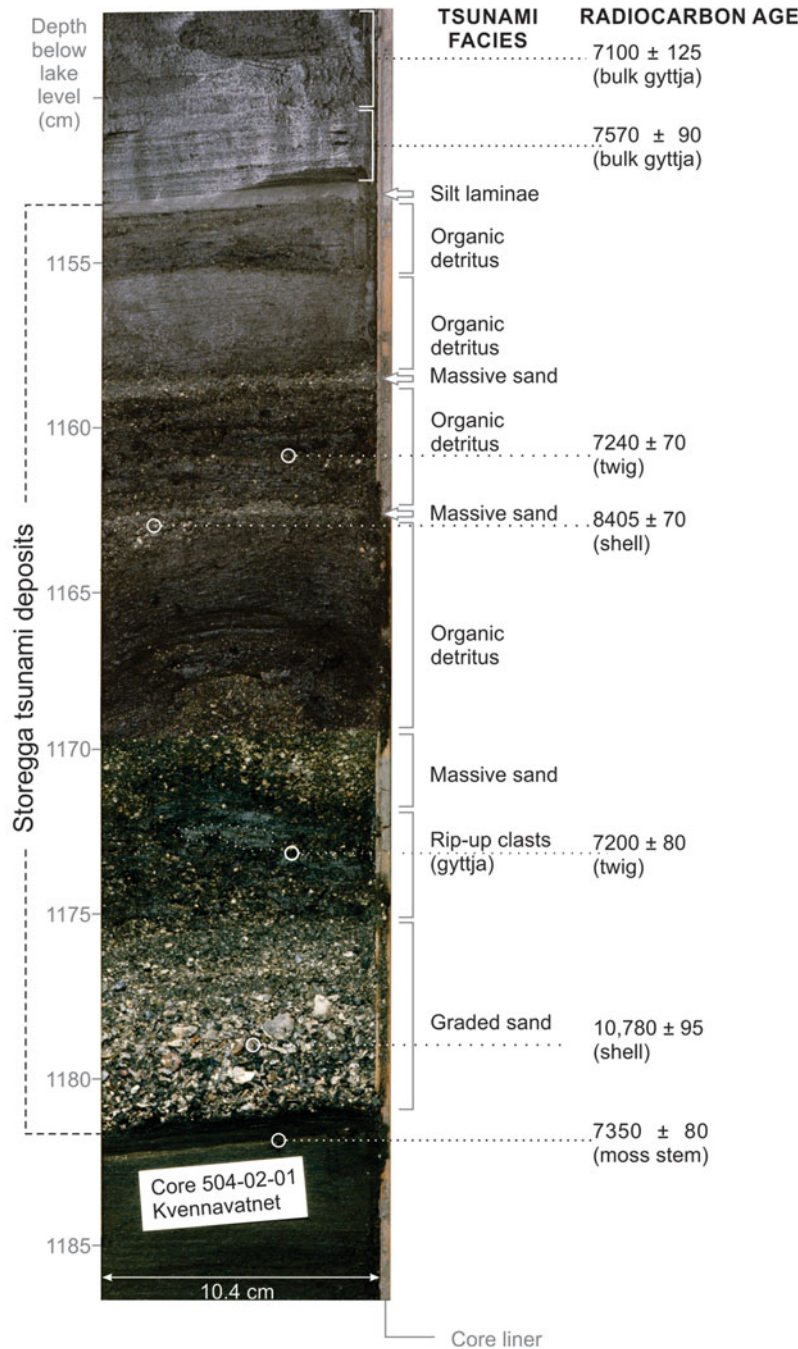
Estuaries in Scotland

In Scotland, the Storegga tsunami deposit is typically a widespread sand layer in estuarine mud (Table 1). An estuary is a drowned valley with an open connection to the sea (Fig. 13). The water in this partly enclosed basin is brackish, because freshwater enters from the upland rivers and mixes with seawater coming in from the ocean. Fine-grained mud accumulates on the floor of the estuary (Fig. 13). Many of the estuaries were formed when glacially scoured valleys were flooded during the sea level rise after deglaciation. Therefore, it is usually peat underneath the

estuarine mud deposits that accumulated when sea level was lower (Fig. 14).

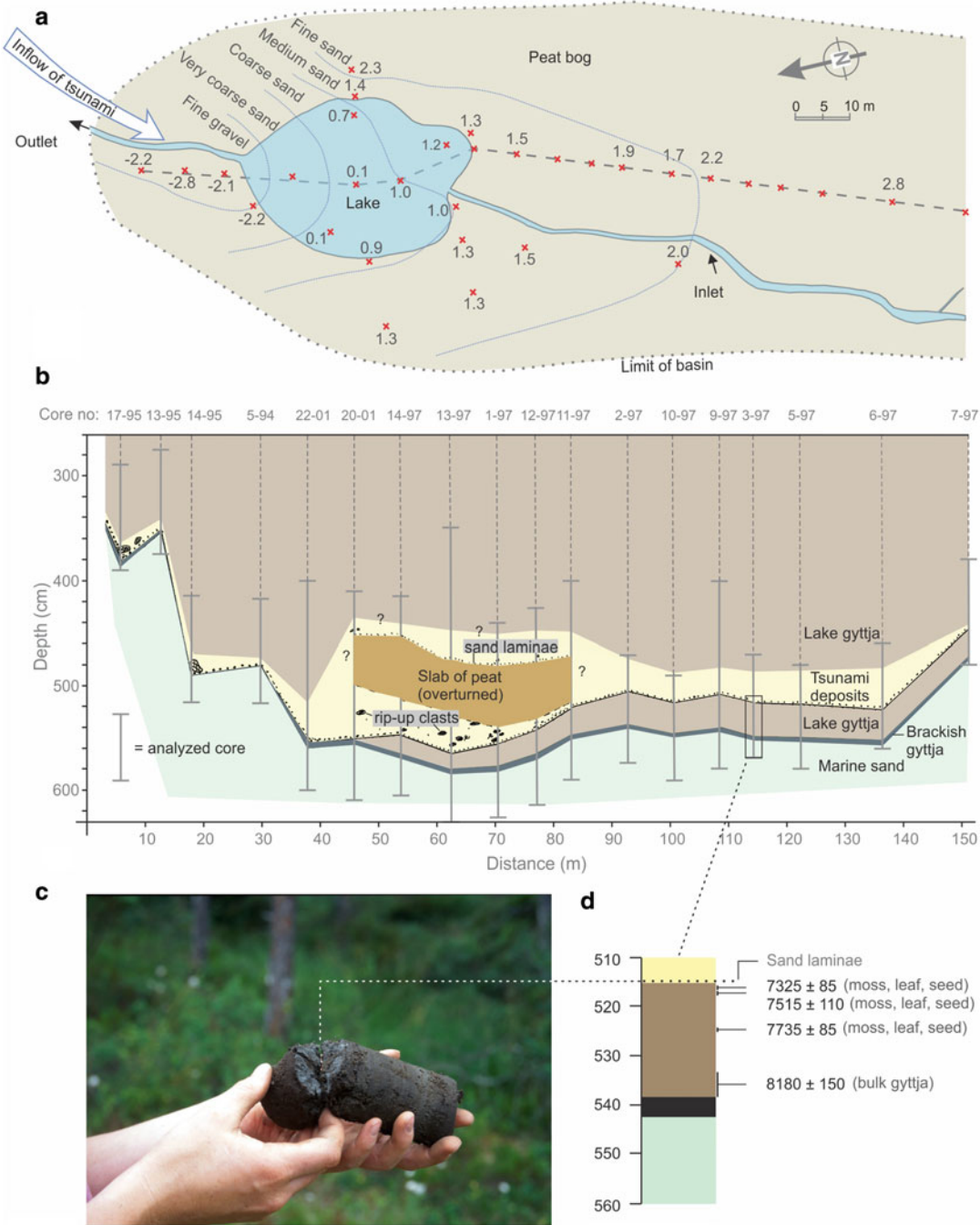
Geologists have successfully used deposits in estuaries to reconstruct sea level changes. The change from peat to overlying mud documents a sea level rise, and a change from mud to overlying peat documents a sea level fall. As sea level rose, the peat/mud boundary moved inland, and when sea level fell, the same boundary moved seaward. By radiocarbon dating the peat just underneath or just above the estuarine mud, a time for when the sea level stood at this elevation, is found. Usually the estuarine mudflats approximate the mean high-water mark of spring tides (Smith et al. 2004). For instance, at Dornoch Firth the modern estuarine mudflats lie 0.3–0.6 m above mean tide level (Smith et al. 1992). It was during studies of estuarine deposits

Tsunami from the Storegga Landslide,
Fig. 11 Tsunami deposits in Kvennavatnet, Bjugn (see location of core in Fig. 9). Depth is cm below lake level. To the right is description of deposits and radiocarbon ages (in ¹⁴C year BP)



to reconstruct sea level changes in Scotland that geologists encountered a widespread sand layer that is now ascribed to the Storegga tsunami event (see section “[How the Storegga Tsunami Was Discovered](#)”).

The Storegga sand layer extends for several hundred meters in the estuary deposits and some places continuous upslope into peat. The layer varies in thickness. In a few places, it is more than 0.5 m thick, for example, in the

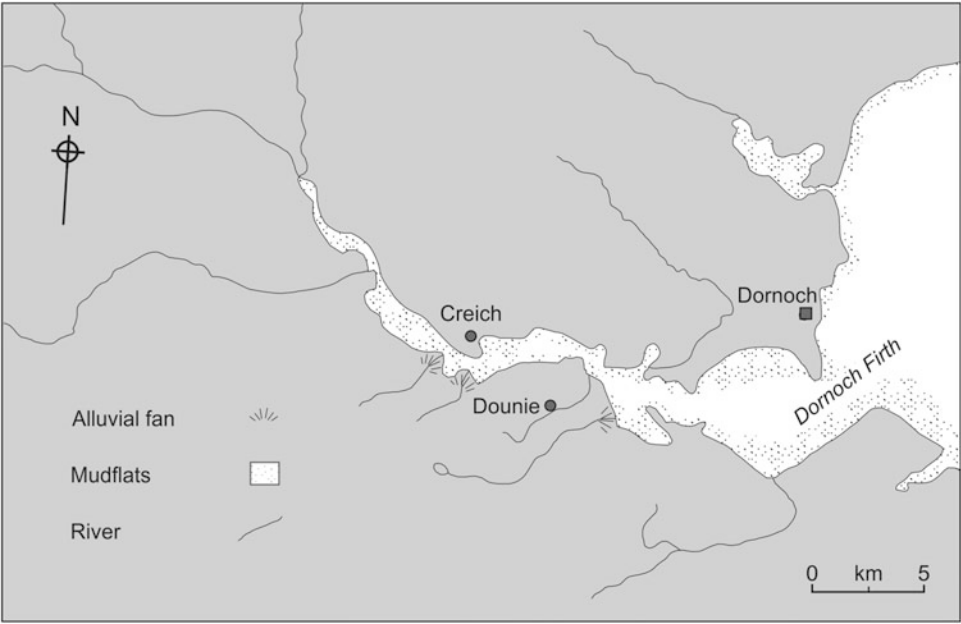


Tsunami from the Storegga Landslide, Fig. 12 Map, cross section, photograph, and log from Gorttjøna I, Bjugn (see location in Fig. 9; slightly modified from Fig. S3 in Bondevik et al. 2012). (a) Map with core locations (red crosses) and grain size in ϕ -units of tsunami sand on the erosional boundary. (b) Cross section along the dashed line in (a). Note the big slab of peat in the tsunami deposits. According to the cores, the peat

slab must be overturned (upside down) because it is more humified at the top than at the bottom. (c) Photo shows the sand lamina on the erosional boundary in core 3-97 – only a few sand grains thick. (d) Radiocarbon ages from core 3 to 97, which was chosen for radiocarbon dating because of minimal erosion underneath the tsunami deposits

Tsunami from the Storegga Landslide, Table 3 Storegga tsunami facies in coastal lakes

Facies	Description
Graded sand	Usually, the first sediment to be deposited on the erosive surface is a graded sand bed (Figs. 8, 10, and 11). The lower part is coarse sand, in some cases fine gravel, grading upward to medium sand. It is commonly rich in shell fragments. Normally 4–8 cm thick, but it may reach a thickness of about 20 cm
Massive sand	No internal structures, poorly to well sorted, from fine gravel to fine sand. It may contain shell fragments. The massive sand bed is normally thinner than the graded sand and varies from 1 mm (one-grain-size thick) (Fig. 12c) to 4–5 cm (Figs. 10 and 11)
Rip-up clasts	We introduced the term organic conglomerate for this facies in Bondevik et al. (1997b) because it resembled a conglomerate; clasts of organic material were distributed in a matrix of lake mud, organic detritus, and sand. The clasts have an irregular form, 0.5–6 cm across (Fig. 11), but may also be much larger (Fig. 12b)
Organic detritus	A mixture of lake mud, plant fragments, and sand without rip-up clasts. It also, usually, is graded, especially if it is present near the top of the tsunami deposits (Fig. 11)
Silt	Some of the tsunami deposits are draped with distinct laminae of silt, 2–3 mm thick (Fig. 11). The silt fines upward from very fine sand or coarse silt to silt. Sometimes the silt can be thicker, 1–2 cm
Fine laminae	In a few lakes, we have seen two-three black laminae in the organic lake mud on top of the tsunami deposits. We think they represent a period of saline bottom water caught in the lake basin in the few years after the tsunami event. This saline bottom water would prevent bioturbation

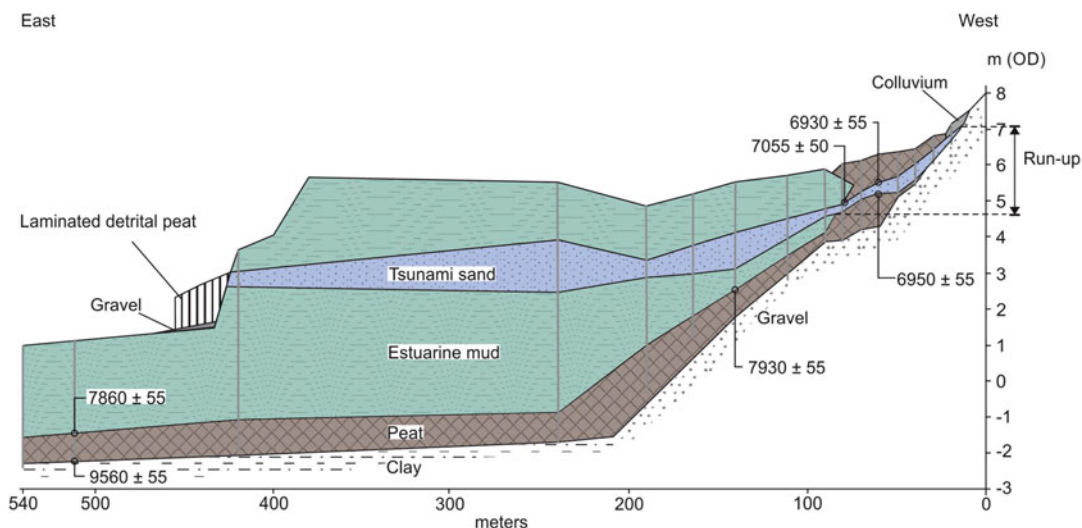


Tsunami from the Storegga Landslide, Fig. 13 Estuary at Dornoch Firth in eastern Scotland. Boreholes at Creich and Dounie (site q in Fig. 1 and

Table 1) show Storegga tsunami deposits in estuarine mud; see profile from Creich in Fig. 14. (Redrawn from Fig. 1 in Smith et al. 1992)

profile in Fig. 14, it is 1.56 m thick in a core ca. 235 m seaward of the limit of the sand. However, in most cores it is between 10 and 30 cm thick. Usually the sand layer tapers off inland and it also fines in grain size in this direction.

Grain-size distribution through the Storegga sand layer, from bottom to top, shows one or two fining-upward sequences. Most samples show a peak in the fine sand fraction, and all sites (except one) register one or two fining-upward sequences (Smith et al. 2004). A few



Tsunami from the Storegga Landslide, Fig. 14 Profile at Creich, cores in gray lines, and ages in radiocarbon years. Run-up is measured from the highest recorded surface of estuarine mud beneath the tsunami

deposit and to the upper reach of the tsunami layer, here measured to 2.3 m. (Redrawn from Long et al. 1989a and Smith et al. 1992)

sites show more than two fining-upward sequences (site **o**; a lagoon in Dawson and Smith (2000) and site **q**; Dawson et al. (1991)). According to Smith et al. (2004), the more sheltered and inland boreholes would have one fining-upward sequence, while more seaward boreholes could have multiple fining-upward sequences. This could mean that the sequences preserved reflect the number of waves inundating the site. Only the highest locations would be inundated by the largest wave and so exhibit one fining-upward sequence.

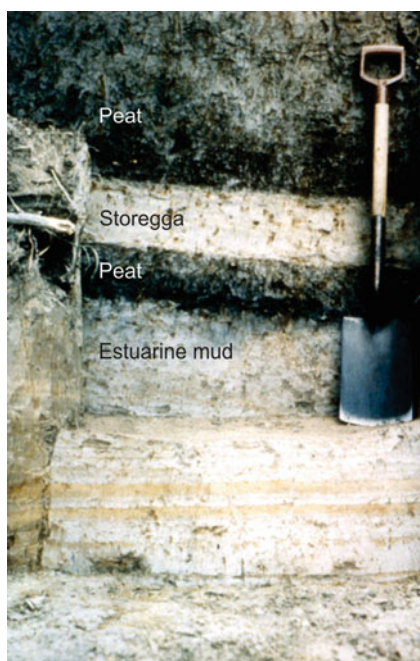
The Storegga sand layer in estuarine settings shows little evidence of erosion (Fig. 15). Rip-up clasts, commonly found in the Norwegian lakes and in Shetland peats, are usually not present. Site **o**, in Sutherland (Fig. 1 and Table 1), has rip-up clasts, but according to its enclosing deposits, it is not an estuary but a lagoon (Dawson and Smith 2000), and its deposits are similar to what we find in coastal lakes in Norway. The lower surface of the tsunami sand layer is usually sharp but shows little variation between boreholes in the former estuaries that could be attributed to erosion.

Run-up of the Storegga sand layer out of the estuary and into former gullies or minor valleys

near the head of the estuaries is between 2 and 5 m (Fig. 1). The run-up is measured from the *highest* altitude of the estuarine mud below the Storegga sand and up to the *highest* altitude of the Storegga sand layer (see Fig. 14). The run-up seems to decrease toward the south along the Scottish coast (Fig. 1).

Peat Outcrops on Shetland

Storegga tsunami deposits are only accessible through coring at most of the sites shown in Fig. 1, but natural outcrops, where the deposits are exposed in a larger section, have been found on the Shetland Islands (site **m** in Fig. 1 and Table 1; Fig. 16). In the Sullom Voe area, there are a few natural outcrops in peat with Storegga tsunami deposits (Bondevik et al. 2003, 2005b; Smith et al. 2004). The outcrops are formed by erosion of storm waves along the shoreline or rivers that cut through the peat surface. Natural outcrops with Storegga layers have not been reported from Norway – a digging machine was used at Harøya (site **f**) to expose peat with Storegga sand underneath a beach ridge (Bondevik 2003b). In Scotland, the only site that



Tsunami from the Storegga Landslide, Fig. 15 Outcrop at Maryton in the Montrose Basin (site t in Fig. 1 and Table 1) shows the Storegga tsunami as a 25 cm thick silty sand deposit between peats. The estuarine mud is laminated silt and clay. (Photo David Smith)

exposes Storegga layers is the cliff at Maryton (Fig. 15).

In this section, I will present a fascinating outcrop in Shetland, near Maggie Kettle's Loch on the western side of Sullom Voe (Fig. 16), which gave us new understanding – in particular about run-up and deposition and formation of rip-up clasts. Below follows a clipping (text in *italic*) from the three paragraphs in Bondevik et al. (2003) that describes the deposits. I have changed the figure number to conform with this paper:

Close to the present shore, the tsunami deposit is 30–40 cm thick and shows large rip-up clasts of peat embedded in the sand (Figs. 17, 18a). Many of the clasts are 10–30 cm in diameter with sharp edges. Also, pieces of wood and trunks were found in the sand. The sand, which is medium to very coarse, contains pebbles and cobbles; we even found a boulder as large as 25 cm in diameter (Fig. 17). The sand thins and fines inland; also, the erosion of peat decreases in this direction

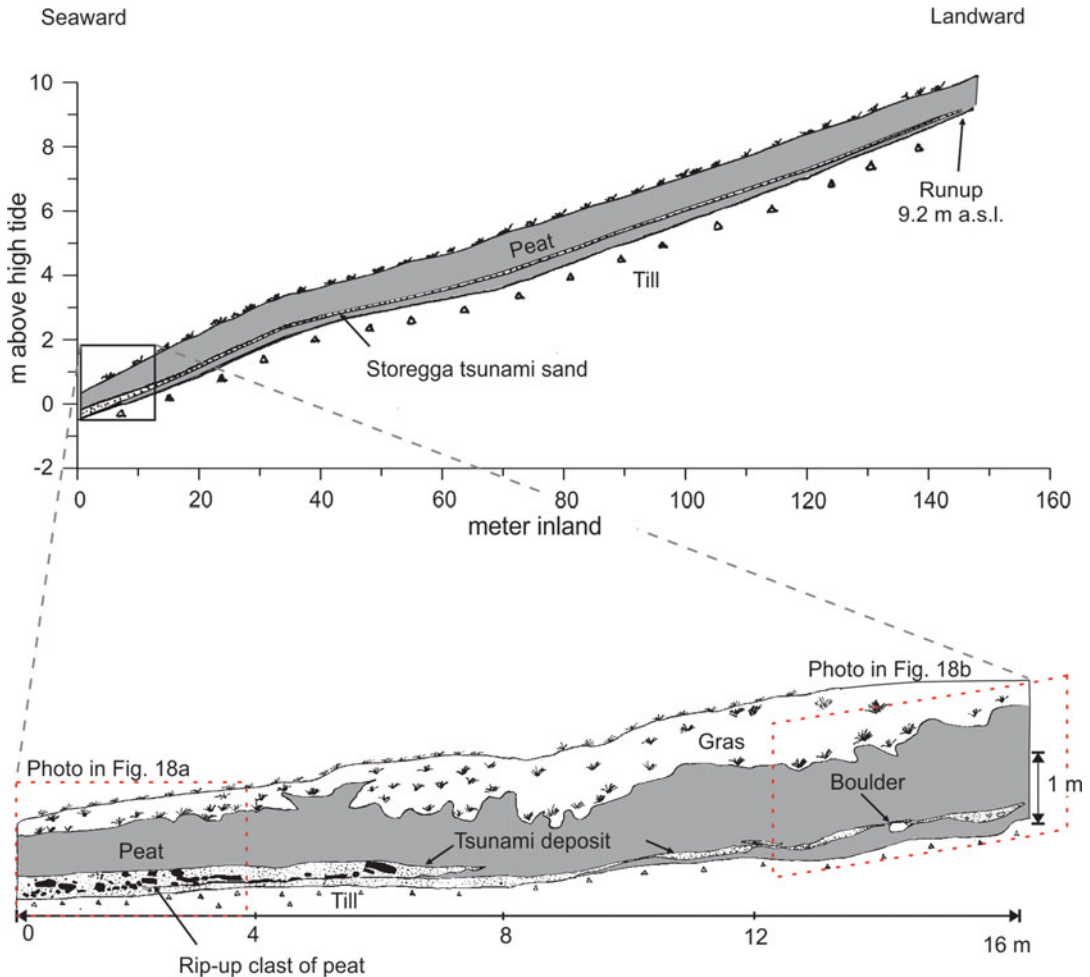


Tsunami from the Storegga Landslide, Fig. 16 Locations of Storegga tsunami deposits on the Shetland Islands. One of the outcrops, indicated by an arrow, in Sullom Voe, is presented in more detail in text and in Figs. 17 and 18

(Figs. 17, 18b). Close to the sea, the sand is 30–40 cm thick. From about 18 m from the shore and inland, the sand thins from 10 cm to less than 1 cm at the maximum elevation (Fig. 17).

Between 0.8 and 4 m above high tide, the sand is normal graded, from very coarse sand with fine gravel particles at the bottom, to medium sand at the top. From 6 m above high tide and inland, the sand is massive—between 4 and 1 cm thick—and discontinuous, and it ends 9.2 m above high tide (Fig. 17).

Rip-up peat clasts, typical for the section between 0 and 6 m, make up a bed within the



Tsunami from the Storegga Landslide, Fig. 17 The upper panel is a sketch of the Storegga tsunami layer in the 150-m-long peat outcrop on the western shore of Sullom Voe (Fig. 16). The lower panel shows the first 16 m of the same outcrop. Between 0 and 6 m, large rip-up clasts of peat and pieces of wood embedded in the sand dominate

the tsunami layer. Underneath the sand layer, there is a sharp erosional unconformity. Here the sand rests on till. From 9 m and inland, the sand is found in peat. Note that the sand layer thickens in the small depressions in the peat. (Redrawn from Fig. 2 in Bondevik et al. 2003)

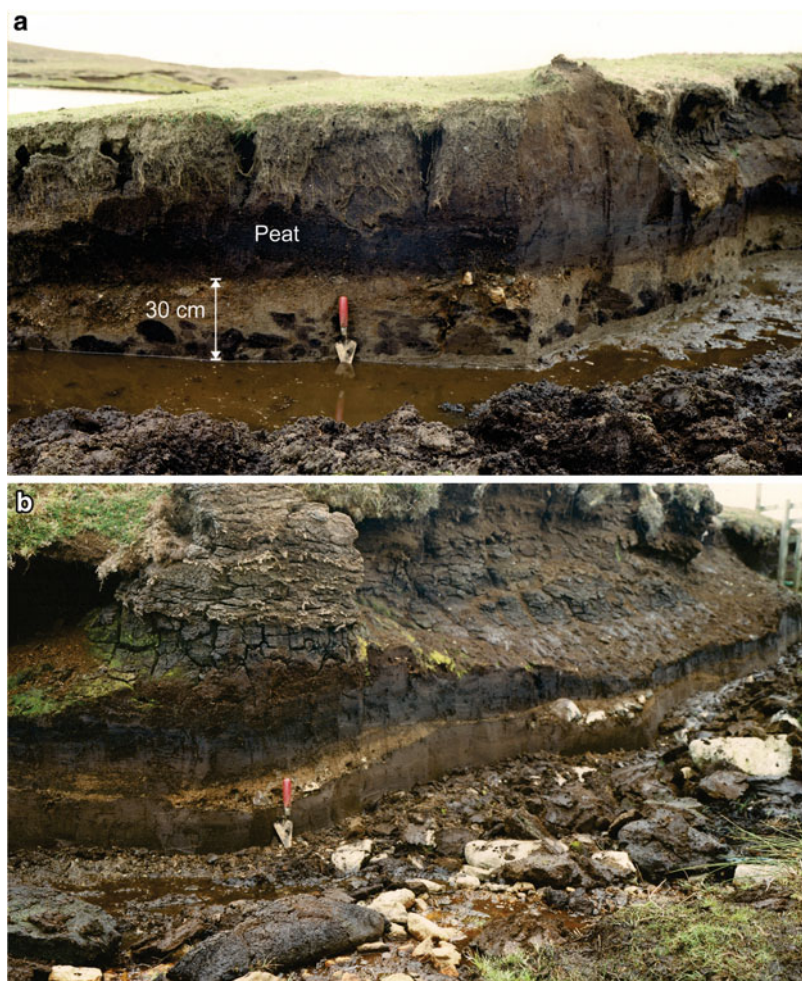
sand, with a distinct lower boundary (Figs. 17 and 18a). We interpret this as a result of at least two waves inundating the land. The first wave eroded the peat surface and transported rip-up clasts of peat and sand. The backwash left the eroded clasts and other organic remains at the surface of the tsunami-laid sand. The following wave buried the clasts in sand.

It is difficult to estimate the run-up on Shetland because we do not know exactly how many meters the sea level at Storegga time was below

present sea level. Hoppe (1965) radiocarbon dated peat found between 8.6 and 8.9 m below the high tide level to between 5990 and 7900 years BP; thus sea level at 6000–7000 years BP was at least 9 m below present sea level. Another point on the sea level curve comes from a marine basin 2 m below present high tide level that prior to 3500 years ago was a freshwater lake (Bondevik et al. 2005b). When the Storegga tsunami happened, relative sea level was clearly lower than 10 m below present sea level. A sea level curve

Tsunami from the Storegga Landslide,

Fig. 18 (a) The first 4 m of the outcrop (location indicated with a red-dashed frame in Fig. 17) shows a large number of rip-up clasts embedded in the sand. Some of the clasts have sharp edges. The lower boundary of the peaty clasts forms a well-defined line in the sand. The red handle on the shovel is 9 cm long. (b) Photo showing the Storegga sand layer from about 12 to 20 m in the peat outcrop (location indicated with another red-dashed frame in Fig. 17). Note how the sand and gravel is thicker in the depressions in the peat. The shovel rests on hard till



constrained by the abovementioned radiocarbon dates indicates that the sea level could be as low as somewhere between 15 m and 30 m below present-day sea level at Storegga time.

Up until now, Shetland has the highest recorded run-up from field evidence. The sand bed in peat at the western shores of Sullom Voe (Fig. 16) was traced to 9.2 m above high tide (Fig. 17). We also traced it in hand cores to continue 2.4 m below high tide in peat underneath the present-day beach gravel. Thus, we have measured a minimum run-up of 11.6 m (Bondevik et al. 2003). In boreholes at Scatsta, on the eastern side of Sullom Voe (Fig. 16), the sand was traced even higher, to 11.8 m above spring tide (Smith et al. 2004). A conservative estimate of the

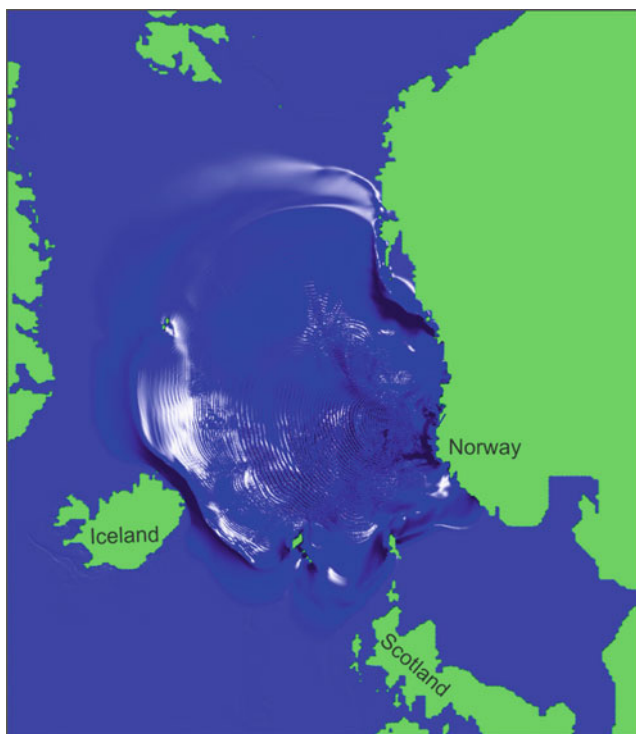
relative sea level when the Storegga tsunami happened is 10–15 m below the present, giving a run-up of more than 20–25 m for the Sullom Voe area in Shetland. Such a high run-up is twice as high as the simulated run-up from the numerical tsunami models.

Numerical Simulations of the Storegga Tsunami

Numerical simulations show how the Storegga tsunami propagated into the North Sea and Norwegian Sea (Fig. 19; <https://doi.org/10.6084/m9.figshare.918635>). In the earlier simulations, the slide was modeled as a box of material that

Tsunami from the Storegga Landslide,

Fig. 19 Simulation of the Storegga tsunami 2 h after the onset of the slide. The wave front, ca. 3 m high, has just reached the Faeroe Islands and Shetland. The small ripples behind the wave front are numerical noise. (Copy of fig. 8 in Bondevik et al. 2005a)



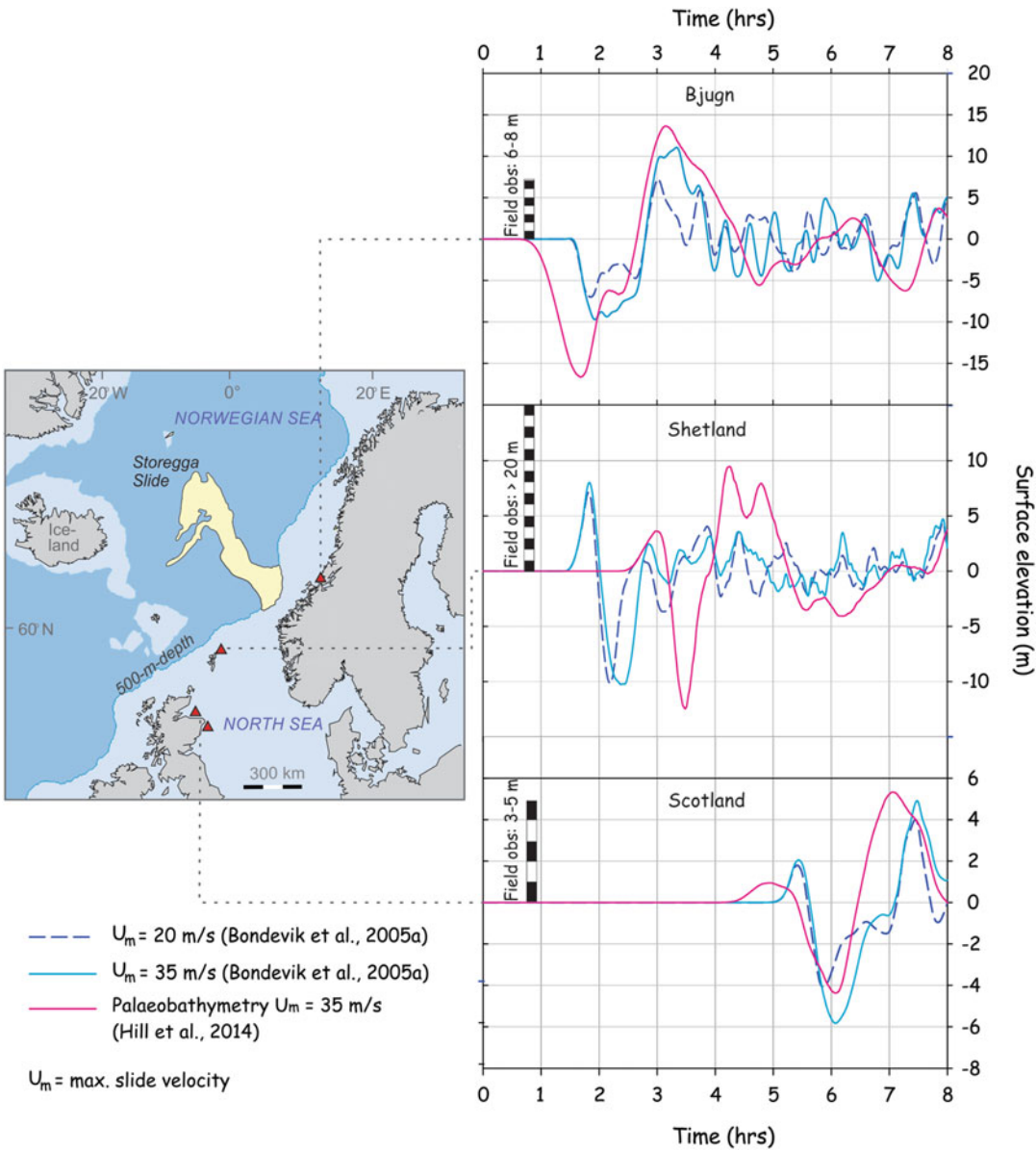
was released at the top of the slope, accelerated over a certain length up to a maximum velocity, and then decelerated until it stopped (Bondevik et al. 2005a; Harbitz 1992; Hill et al. 2014). In a recent paper, Løvholt et al. (2017) use more realistic slide movements. Their first model has a retrogressive slide motion (see section “The Slide that Triggered the Tsunami”) where the slide fails as a continuous sequence of blocks that started in deeper water and retreated up the slope. The other simulation models the slide as a debris flow. Overall, the different models return run-up heights in agreement with the field observations.

The modeled slides attempt to match the reconstructed morphology in the source area of the slide. Harbitz (1992) and Hill et al. (2014) modeled the slide as a box with a near-uniform thickness of 144 m, width 85 (Harbitz 1992) and 175 km (Hill et al. 2014), and a length of 150 km. The slide box in Bondevik et al. (2005a) represents the excavated area better as it varies in thickness according to the pre-slide bathymetry. This block has a maximum thickness of 400 m at

the headwall and tapers off to zero at the end (length 150 km) with a volume of 2400 km³. The debris flow model of Løvholt et al. (2017) involves a volume of ~3000 km³.

Maximum frontal velocity of the slide is 30–35 m/s in the different simulations. This velocity comes from two sources: the observations from the 1929 Grand Banks Slide and numerical simulations of the slide movement itself (De Blasio et al. 2005; Løvholt et al. 2017). Telephone cables between Europe and America broke subsequently as the Grand Banks Slide propagated downslope. According to the different cable breaks, the slide velocity was estimated to 28 m/s at about 150–200 km from the shelf edge (Heezen and Ewing 1952). A maximum frontal speed of 35 m/s for the slide is reasonable, and it generated tsunami waves that fitted rather well with the field observations of run-up (see below). Bondevik et al. (2005a) also simulated the tsunami using a lower maximum slide velocity of 20 m/s (Fig. 20).

In the slide box simulations, the slide accelerates half the runout distance (75 km) where it



Tsunami from the Storegga Landslide, Fig. 20 Simulated sea surface elevation at Bjugn, Shetland, and Scotland during the Storegga tsunami. The locations are identical for both simulations, except Scotland where Hill et al. (2014) has the northernmost location. The

two simulations differ in arrival time of the tsunami because they use different bathymetry. Hill et al. (2014) have higher resolution of the bathymetry near the coastline of Scotland and Shetland, which also is corrected for changes in sea level the last 8000 years

reaches a velocity of 35 m/s (after 56 min) and then immediately starts to decelerate for another 75 km until it stops at a runout length of 150 km. The slide box moves for about 1.9 h (2×56 min). The runout length of 150 km is close to where the

real slide disintegrated from a debris flow into turbidity currents that ran farther down the slope (Fig. 2).

In the numerical simulations, the Norwegian Sea/North Sea with coastlines is divided into

polygons (often squares), called a grid or a mesh, with a certain resolution (fixed to, e.g., 500×500 m, or variable) and a set of equations that calculate the flow of water in and out of the grid cells (boxes). The equations used are either linear shallow water equations (Bondevik et al. 2005a; Harbitz 1992; Løvholt et al. 2017) or Navier-Stokes equations (Hill et al. 2014). Both models are derived from the assumptions of conservation of mass and conservation of linear momentum. If more water flows into a grid cell than flows out of the grid cell, the water level in that cell must rise because water is not compressible (conservation of mass). The other equations are derived from Newton's second law, written for fluid motion under the assumption that the momentum is conserved.

The simulations show that the tsunami propagates outward in all directions from the slide area. A large depression of 5–8 m moves toward the Norwegian coast as the slide moves downslope, whereas a positive wave of about 3 m in height moves seaward toward Shetland, Iceland, and Greenland – most of the energy is transferred in the same direction as the slide moves (Fig. 19). About 30 min after the slide was released, the first and negative wave reaches the Norwegian coast. From 2 h (Bondevik et al. 2005a) to 3 h (Hill et al. 2014) after the slide started, the first wave reached Shetland and the Faeroe Islands (Fig. 20).

Over all, the simulated waves in all models compare rather well with the field observations. The box models tend to overestimate the run-up along the western coast of Norway by 20–40% (Fig. 20) (Bondevik et al. 2005a; Hill et al. 2014), but on the Faeroe Islands and Shetland Islands, the run-up of deposits is more than 50% higher than the largest simulated waves (Fig. 20). Here the observations are from within fjords and sounds (Fig. 16), whereas the simulations are from the very coast (Fig. 20). We think the larger observed run-up reflects amplification within the narrowing and shallowing fjords. The simulations of Hill et al. (2014) gave a slightly better fit to the sediment run-up in some areas probably because they used a more detailed bathymetry of the coastlines and a bathymetry that included changes in sea

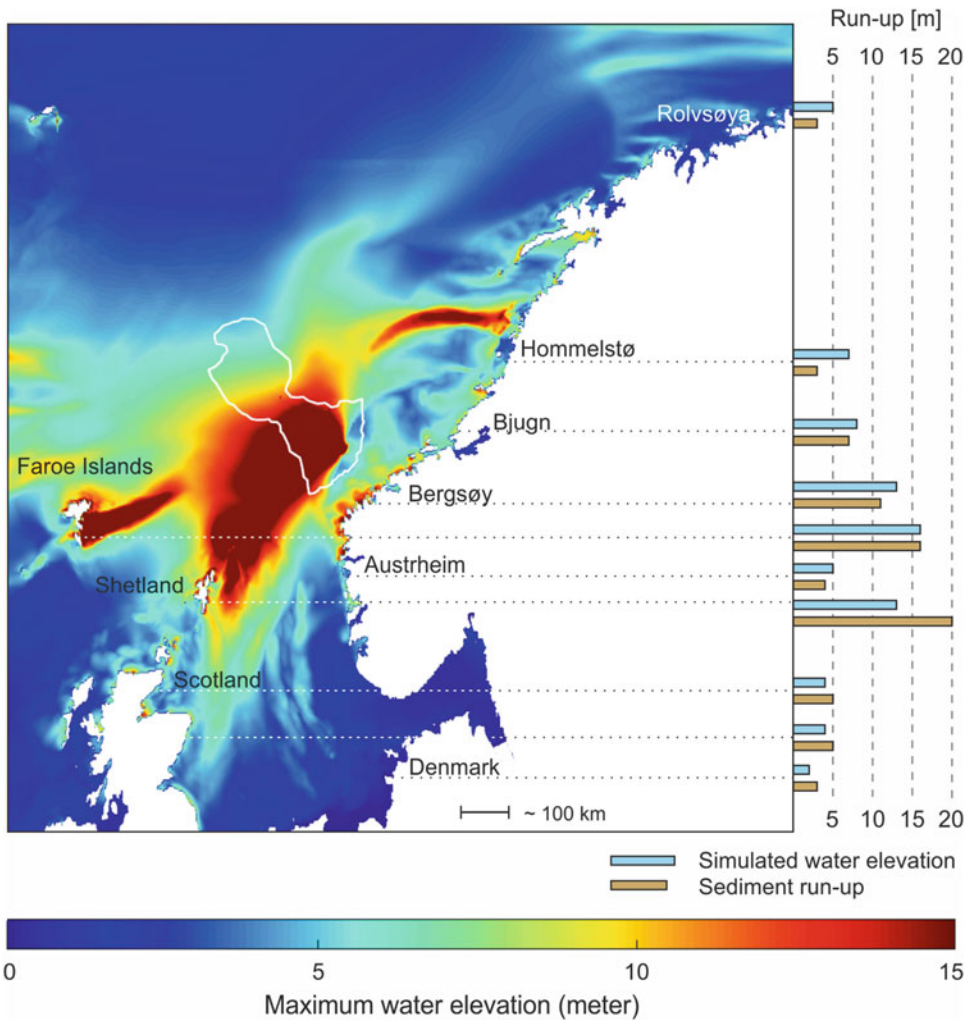
level since 8000 years ago. Of the simulations, the debris flow model of Løvholt et al. (2017) produces surface elevations that matches the observations the best (Fig. 21).

Maximum wave height – or more precisely called maximum surface elevation – depends on the volume and movement of the sliding masses, as well as the water depth in the slide area (Harbitz et al. 2006; Løvholt et al. 2005). It is important to remember that as the slide moves it forms a bulge on the sea surface in the slide direction and a depression at the rear – both propagates at a velocity of $c = \sqrt{gh}$, g is acceleration of gravity, and h is water depth. For submarine slides the propagating tsunami waves usually move faster than the slide and that limits the buildup of the wave. The maximum surface elevation depends on the thickness of the slide, the velocity and acceleration of the slide, and the propagating velocity of the tsunami waves (which depends on water depth). For simulated landslides in the Storegga escarpment, Løvholt et al. (2005) found that both the products of initial acceleration and volume, $a_0 \times V$, and maximum velocity and volume, $U_m \times V$, correlate well with the maximum surface elevation of the generated tsunami.

Dating the Storegga Event

Tsunami deposits are often difficult to date accurately. The radiocarbon clock starts when an organism dies, and the challenge has been to find the remains of plants or animals that were actually killed in the tsunami and not just remains of already dead organisms that have been redeposited. Only radiocarbon ages of the true victims will return an accurate age of when the tsunami happened. However, most of the organic material within a tsunami deposit is redeposited and older than the tsunami event itself and comes from the heavy erosion by the tsunami of older deposits along the shores containing “dead” organic material (e.g., Jankaew et al. 2008).

From the beginning, the Storegga tsunami deposits were dated on samples of bulk organic material resting directly upon/or below the



Tsunami from the Storegga Landslide, Fig. 21 Maximum sea surface elevations simulated with a debris flow model. The white outline shows the observed runout of the landslide debris. The turbidities continued

farther (see Fig. 2). To the right is a comparison of maximum simulated sea surface elevations and sediment run-up. (Figure redrawn from Fig. 3 in Løvholt et al. 2017)

Storegga tsunami deposits. Smith et al. (2004) found an age of ca. 7000 ^{14}C year BP from over 50 such dates in Scotland and concluded that there might have been a time delay for the start of peat accumulation on top of the tsunami deposits because the ages were about 200–300 years younger than radiocarbon dates of better and other samples (see below). Another problem arises from the penetration of roots into the deeper layers of peat. Roots can transfer current atmospheric CO_2 -carbon to deeper layers, thus reducing the

radiocarbon age of the peat. It is also likely that roots could grow through the tsunami sand layer and into the peat below. Radiocarbon ages shown in Fig. 14 probably illustrate this possibility.

Another and better way is to date individual fragments from the aboveground parts of plants, like twigs, leaves, fruits, and seeds washed out from the deposits. Based on a careful selection of such fragments, Bondevik et al. (1997a) proposed that Storegga tsunami dates to 7250–7350 ^{14}C year BP (see Fig. 11 with the ages of twigs). In a

comprehensive study of many of the radiocarbon ages ($n = 127$) of the Storegga tsunami, Weninger et al. (2008) concluded the event to have happened at 7300 ± 30 ^{14}C year BP, corresponding to the interval 8200–8000 years BP (2- σ range).

The best material found so far for radiocarbon dating is samples of green moss exceptionally well preserved within the Storegga tsunami deposits. Such moss fragments were found in backwash deposits at site **c** (Fig. 22) and **d** (Fig. 23) in Northern Norway (Fig. 1 and Table 1). The mosses are still green colored because they contain small amounts of chlorophyll. A little of their chlorophyll survived because the tsunami buried the mosses in shell-rich sediments below a protecting layer of marine mud (Fig. 23). These sediments preserved the chlorophyll by keeping out light and oxygen and by keeping the pH above 7. Because of their preserved green color, we know the green mosses were buried alive and their radiocarbon clock started ticking within hours after the Storegga Slide had set off the tsunami (Bondevik et al. 2012).

The green moss species within the Storegga tsunami deposits were radiocarbon dated to a weighted mean of 7300 ± 20 ^{14}C year BP, corresponding to 8030–8180 years BP on a 2- σ level. This mean combines seven ages, each on a

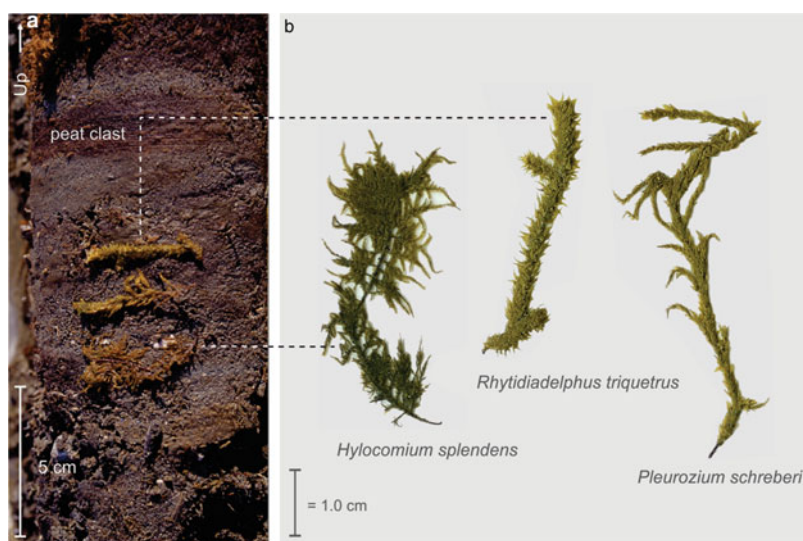
different piece of green moss, with ages ranging from 7231 ± 64 to 7387 ± 72 ^{14}C year BP (Fig. 23). Through Bayesian analysis, also including radiocarbon ages of other samples from below and above the tsunami deposit, the calibrated age interval is narrowed to 8120–8175 (1- σ level) and to 8070–8180 (2- σ level) years BP, with 8150 as the most probable year for the Storegga event (for details see Bondevik et al. 2012).

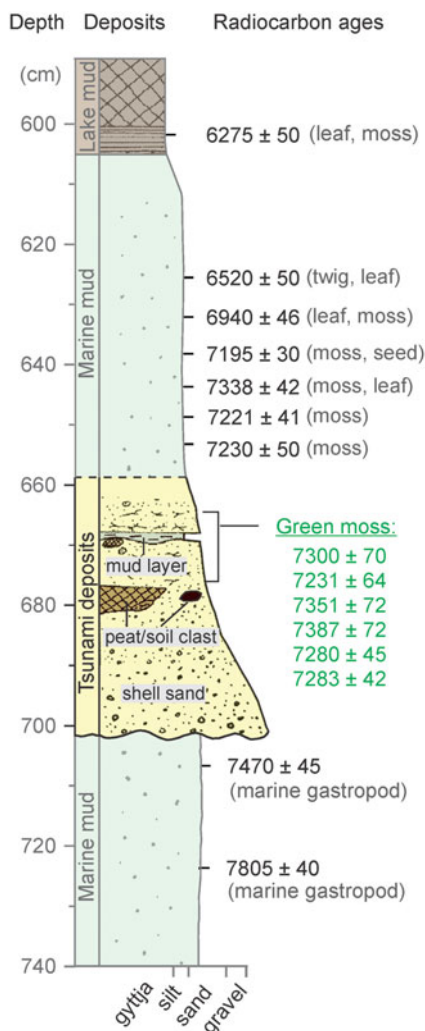
It was also a challenge to obtain accurate ages of the Storegga Slide itself. The slide cuts deep into older layers, and the slide material is thus much older than the event itself (Figs. 2, 3, 4, and 5). Most dates come from radiocarbon ages of foraminifera that accumulated in mud on top of the slide deposits after the slide event. Those ages are thus younger than the event. Based on many radiocarbon ages, Hafliðason et al. (2005) concluded that the age of the slide is 8200 ± 250 years BP (Table 2) – in agreement with the date of the tsunami deposits.

The green mosses also told another story – that they were killed in the fall, possibly in October (Rydgren and Bondevik 2015). One of the species of the green mosses was *Hylocomium splendens* that grows in a regular pattern. New segments on the moss, called daughters, branch off from the previous year's segments, called the mother segments (Fig. 24). The size ratio between the

Tsunami from the Storegga Landslide,

Fig. 22 Samples of green moss in Storegga tsunami deposits in Lyngen, Troms (site **c**, Fig. 1). To the left (**a**) is a photo of the core with tsunami backwash sediments. Terrestrial moss samples, washed out from a layer of shell fragments in the core, are placed clean on the core surface for display. The same moss samples to the right (**b**) after being dried in the laboratory – the green color of the moss samples shows much better after drying





Tsunami from the Storegga Landslide, Fig. 23 Storegga tsunami deposits from core site 5 at Djupmyra in Hommelstø (site **d** in Fig. 1). Uncorrected radiocarbon ages are given to the right. Weighted average of the green moss dates is 7320 ± 20 ^{14}C year BP, calibrated to 8070–8180 years BP (Bondevik et al. 2012). Note that the tsunami deposits contain backwash deposits of peat and soil clasts and terrestrial plants, preserved beneath a cover of marine mud (copy of Fig. 2 in Rydgren and Bondevik 2015)

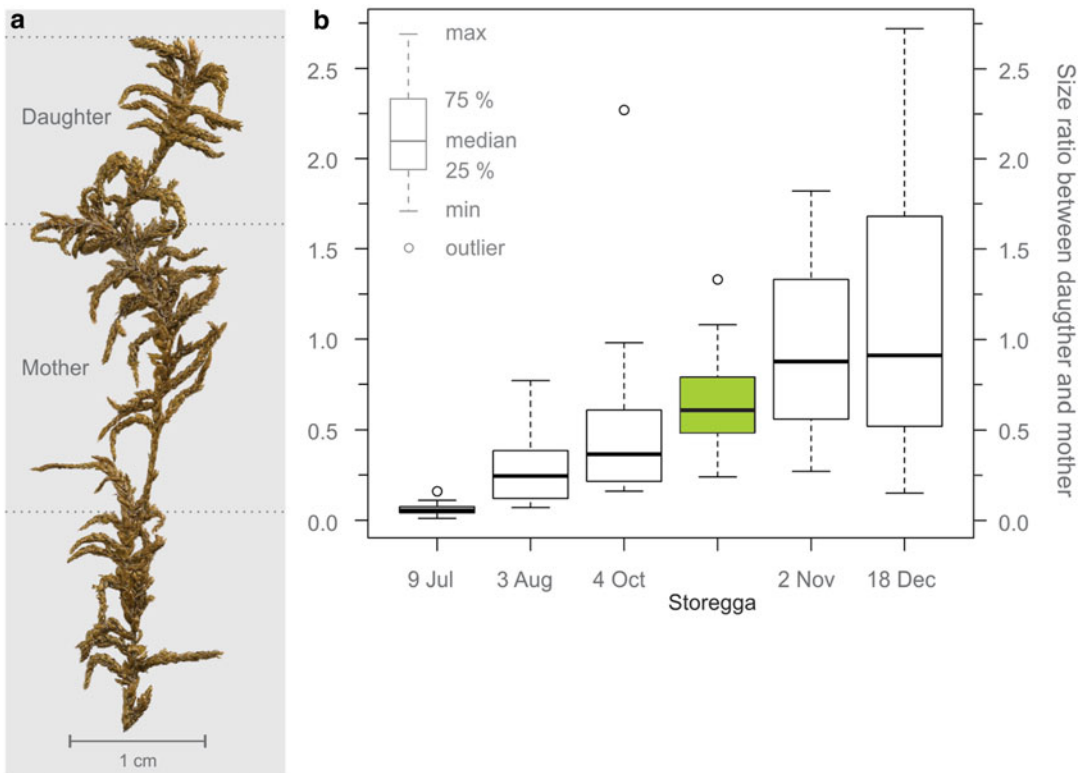
daughter and the mother segments indicates the time of the year. From measuring the size ratio between the daughter and the mother segments of 19 samples of *Hylocomium splendens*, Rydgren and Bondevik (2015) were able to suggest that the Storegga mosses were killed in

October–November (Fig. 24). The giant tsunami happened late in the fall – so late in the year that humans living along these coasts were most likely near their seashore settlements preparing for the coming winter – and not in the mountains to hunt reindeer.

Storegga Tsunami and Stone Age Humans

Mesolithic humans occupied the coasts of Norway and Scotland and must have been hit by the Storegga tsunami. Remains of Mesolithic settlements are often located on or near the former shores – thus their settlements would be very vulnerable to a tsunami. Rock carvings illustrate hunting and fishing from boats, and settlements on distant islands suggest that they had advanced maritime skills at that time (Bjerck 2013). However, evidence that prove that humans or settlements were actually hit by the Storegga tsunami have not yet been discovered, although archeologists have been aware of the Storegga tsunami for more than a decade. Sands of Storegga age cover two Mesolithic settlements, at Dysvikja in Western Norway (near site **f**, Fig. 1) and Inverness in Scotland (site **r** in Table 1, Fig. 1), but we do not know if humans still occupied these settlements when the Storegga tsunami hit or if the humans had already left the sites for other reasons, like the rising sea level during the mid-Holocene transgression (Tapes transgression).

The same might be true for Doggerland (Coles 1998), a low-lying island in the North Sea, today about 20 m below sea level, occupied by humans in Late Paleolithic and Early Mesolithic. Hill et al. (2014) used ocean depths in their tsunami model that were corrected for changes in sea level that have occurred the last 8000 years, and at that time Doggerland was just above sea level by a few meters. In their simulation, you can see how Doggerland is hit by the tsunami (<https://doi.org/10.6084/m9.figshare.918635>); the largest wave is about 5 m. However, the question is whether Doggerland was still occupied at that time or whether humans had already left the island (s) because of the rising sea level (<http://www>.



Tsunami from the Storegga Landslide, Fig. 24 (a) A well-preserved sample of the moss *Hylocomium splendens* from Storegga tsunami deposits at Djupmyra, Hommelstø (Site d, Figs. 1 and 23). The daughter segment is a little smaller than the mother segment. For this sample the daughter/mother size ratio is 0.63. (b) Plot of the size

ratio between daughter and mother segments of *Hylocomium splendens* from modern samples ($n = 20$) collected in the months July, August, October, November, and December. The size ratio of Storegga samples ($n = 19$) is somewhere between 4 October and 2 November. (From Fig. 4 in Rydgren and Bondevik 2015)

bbc.com/news/science-environment-27224243). The youngest radiocarbon ages of human remains and artifacts from Doggerland retrieved so far are all older than Storegga time (Weninger et al. 2008). This would suggest that Doggerland was probably already abandoned by the time of the Storegga event.

At Dysvikja at Fjørtoft (a neighbor Island to site f in Fig. 1; Table 1), a Mesolithic settlement of Storegga age was discovered below 1.2 m of beach gravel. The site contained flint tools, hazelnuts, three fireplaces, birch bark, and other materials possibly part of a building construction (Indrelid 1973). The critical point is a sand layer, between 7 and 15 cm thick, found above the cultural beds dated to 7550 ± 90 ^{14}C year BP (Indrelid 1974) and below the 1.2 m of overlying

beach ridge (Tapes beach ridge). When discovered, the sand layer was interpreted as deposited from wind, but this interpretation was questioned by Bondevik (2003b), who found a similar stratigraphy (without the archeological material) at site f (Fig. 1), where a sand layer in peat below the Tapes beach ridge was interpreted to be deposited from the Storegga tsunami.

Rydgren and Bondevik (2015) found that Storegga happened in October/November and speculate that autumn must have been a difficult time for the Late Mesolithic humans to be hit by the tsunami. At that time of the year, most of the hunter-gatherer groups adapted to the coastal environment in Western Norway were back at the coast from visits in the mountains to hunt reindeer and/or moose in late summer and early

autumn (Bang-Andersen 1996; Bjerck 2008). A tsunami in late autumn, after the hunters returned to the coast, could have caused high mortality. Based on a decline in the number of radiocarbon ages of archeological remains between 8400 and 7600 years BP in northeast Britain, Waddington and Wicks (2017) speculate that this might reflect a collapse in the population partly caused by the Storegga tsunami. For those who survived, the loss and destruction of dwellings, boats, clothing, equipment, and food supplies would have made the following winter very difficult (Bjerck 2008; Rydgren and Bondevik 2015).

Future Directions

How hard did the Storegga tsunami hit the Stone Age humans? At present, we do not know. Most of the settlements were probably at the coast, and they were most likely severely damaged by the tsunami, but direct evidence have not yet been found, although the archeologists are well aware of the Storegga tsunami event and have looked for it during archeological excavations. When future Mesolithic settlements at the coast of this region are unearthed through excavations, archeologists should have the Storegga tsunami in mind. The two sites Dysvikja and Inverness are covered by Storegga sand, but the sites could have been abandoned before the tsunami happened.

The Storegga tsunami deposit can be a useful stratigraphic marker because of its large extent, short duration, and extensive deposits recognizable in many different coastal settings. The green mosses date the Storegga event to 8070–8180 years BP (2- σ level) (Bondevik et al. 2012), but a tree ring date of the Storegga event could date the event to the nearest decade. A prerequisite is that we must find remains of an old tree with the outer bark preserved that was killed during the tsunami. Some of the lake basins or bogs must have trapped parts of fallen trees during the tsunami. If we could find samples of such trees with some decades of rings, we might be able to wiggle-match the radiocarbon ages and get a better date for the Storegga event.

Are there evidence of other tsunami events than the Storegga in the Norwegian Sea/North Sea? The short answer is yes. On Shetland we found two younger tsunami deposits, at 5500 and 1500 years BP (Bondevik et al. 2005b), but we do not know what triggered these events or how widespread they are. Another big slide north of Storegga is the Traenadjupet Slide that happened about 4500 years BP (Laberg and Vorren 2000). Could that slide also have triggered a tsunami? So far we have not found evidence for that, but that could be because we have searched in wrong places. The numerical simulation of a possible tsunami generated by the Traenadjupet Slide could help us to choose the area along the coast that would have the largest run-up (Løvholt et al. 2017). However, none of these events compares to the giant Storegga. If Storegga happened today, it would have caused a big catastrophe – comparable in size to the 2004 Indian Ocean tsunami.

Acknowledgments I am grateful to the MAREANO project at Norwegian Geological Survey (NGU) and the Norwegian Hydrographic Service that allowed me to use their detailed bathymetry for Figs. 4 and 5, which were nicely put together by Terje Thorsnes at NGU. Finn Løvholt and Jon Hill answered questions about the numerical simulations and also commented on the relevant text in that chapter; David Smith answered detailed questions about run-up measurements in Scotland, provided the photo from Maryton in Fig. 15, and suggested changes that improved the paper. Alastair Dawson and John Inge Svendsen gave valuable comments to the entry “How the Storegga Tsunami Was Discovered”? Robert I. Tilling reviewed the paper.

Bibliography

- Aksdal S (1986) Holocene vegetasjonsutvikling og havnivåendringer i Florø, Sogn og Fjordane. Bergen, Bergen, 105pp
- Bang-Andersen S (1996) Coast inland relations in the Mesolithic of southern Norway. *World Archaeol* 27(3):427–443
- Befring S (1984) Submarine massebevegelser nordvest av Storegga utenfor Møre og Romsdal: En genetisk klassifisering av og en regional oversikt over de øvre deler av sedimentene i rasområdet. Bergen, Bergen, 95pp
- Berg K, Solheim A, Bryn P (2005) The Pleistocene to recent geological development of the Ormen Lange area. *Mar Pet Geol* 22(1–2):45–56

- Bjerck HB (2008) Norwegian Mesolithic trends. In: Bailey GN, Spikins P (eds) *Mesolithic Europe*. Cambridge University Press, Cambridge, pp 60–106
- Bjerck HB (2013) Looking with both eyes. *Nor Archaeol Rev* 46(1):83–87
- Bondevik S (2003a) Records of paleo-tsunamis around the Norwegian Seas. University of Tromsø, Tromsø
- Bondevik S (2003b) Storegga tsunami sand in peat below the Tapes beach ridge at Harøy, western Norway, and its possible relation to an early Stone Age settlement. *Boreas* 32(3):476–483
- Bondevik S, Svendsen JI, Johnsen G, Mangerud J, Kaland PE (1997a) The Storegga tsunami along the Norwegian coast, its age and runup. *Boreas* 26(1):29–53
- Bondevik S, Svendsen JI, Mangerud J (1997b) Tsunami sedimentary facies deposited by the Storegga tsunami in shallow marine basins and coastal lakes, western Norway. *Sedimentology* 44(6):1115–1131
- Bondevik S, Mangerud J, Dawson S, Dawson A, Lohne Ø (2003) Record-breaking height for 8000-year-old tsunami in the North Atlantic. *Eos, Trans Am Geophys Union* 84(31):289
- Bondevik S, Løvholt F, Harbitz C, Mangerud J, Dawson A, Inge Svendsen J (2005a) The Storegga Slide tsunami—comparing field observations with numerical simulations. *Mar Pet Geol* 22(1/2):195–208
- Bondevik S, Mangerud J, Dawson S, Dawson A, Lohne Ø (2005b) Evidence for three North Sea tsunamis at the Shetland Islands between 8000 and 1500 years ago. *Quat Sci Rev* 24(14–15):1757–1775
- Bondevik S, Stormo SK, Skjerdal G (2012) Green mosses date the Storegga tsunami to the chilliest decades of the 8.2 ka cold event. *Quat Sci Rev* 45:1–6
- Boomer I, Waddington C, Stevenson T, Hamilton D (2007) Holocene coastal change and geoarchaeology at Howick, Northumberland, UK. *Holocene* 17(1):89–104
- Bryn P, Berg K, Forsberg CF, Solheim A, Kvalstad TJ (2005a) Explaining the Storegga Slide. *Mar Pet Geol* 22(1–2):11–19
- Bryn P, Berg K, Stoker MS, Haflidason H, Solheim A (2005b) Contourites and their relevance for mass wasting along the Mid-Norwegian Margin. *Mar Pet Geol* 22(1–2):85–96
- Bryn P, Jasinski ME, Søreide F (2007) Ormen Lange – pipelines and shipwrecks. Universitetsforlaget, Oslo, 174pp
- Bugge T (1983) Submarine slides on the Norwegian continental margin, with special emphasis on the Storegga area, vol 110. Continental Shelf Institute Publication, Trondheim, p 152
- Bugge T, Lien R, Rokoengen K (1978) Kartlegging av løsmassene på kontinentalsokkelen utenfor Møre og Trøndelag: lettseismisk profilering (Quaternary deposits off Møre and Trøndelag, Norway: seismic profiling). Continental Shelf Institute Publication, 99:1–55
- Bugge T, Befring S, Belderson RH, Eidvin T, Jansen E, Kenyon NH, Holtedahl H, Sejrup HP (1987) A giant three-stage submarine slide off Norway. *Geo-Mar Lett* 7(4):191–198
- Coles BJ (1998) Doggerland: a speculative survey. *Proc Prehist Soc* 64:45–81
- Corner G, Haugane E (1993) Marine-lacustrine stratigraphy of raised coastal basins and postglacial sea-level changes at Lyngen and Vanna, Troms, northern Norway. *Nor J Geol* 73:175–197
- Dawson S, Smith DE (1997) Holocene relative sea-level changes on the margin of a glacio-isostatically uplifted area: An example from northern Caithness, Scotland. *Holocene* 7(1):59–77
- Dawson S, Smith DE (2000) The sedimentology of Middle Holocene tsunami facies in northern Sutherland, Scotland, UK. *Mar Geol* 170(1–2):69–79
- Dawson AG, Long D, Smith DE (1988) The Storegga Slides: evidence from eastern Scotland for a possible tsunami. *Mar Geol* 82(3–4):271–276
- Dawson AG, Smith DE, Long D (1990) Evidence for a tsunami from a mesolithic site in Inverness, Scotland. *J Archaeol Sci* 17(5):509–512
- Dawson AG, Foster IDL, Shi S, Smith DE, Long D (1991) The identification of tsunami deposits in coastal sediment sequences. *Sci Tsunami Haz* 9:73–82
- De Blasio FV, Elverhøi A, Issler D, Harbitz CB, Bryn P, Lien R (2005) On the dynamics of subaqueous clay rich gravity mass flows—the giant Storegga slide, Norway. *Mar Pet Geol* 22(1–2):179–186
- Edvin T (1984) Stratigrafiske undersøkelser av kjernepøver fra rasområdet utenfor Storegga. Bergen, Bergen, 122pp
- Fægri K (1944) Studies on the Pleistocene of western Norway. III. Bømlo. Bergen Museums Årbok. Naturvit. rk. 1943
- Fruergaard M, Piasecki S, Johannessen PN, Noe-Nygaard-N, Andersen TJ, Pejrup M, Nielsen LH (2015) Tsunami propagation over a wide, shallow continental shelf caused by the Storegga slide, southeastern North Sea, Denmark. *Geology* 43(12):1047–1050
- Grauert M, Björck S, Bondevik S (2001) Storegga tsunami deposits in a coastal lake on Suouroy, the Faroe Islands. *Boreas* 30(4):263–271
- Haflidason H, Sejrup HP, Nygård A, Mienert J, Bryn P, Lien R, Forsberg CF, Berg K, Masson D (2004) The Storegga Slide: architecture, geometry and slide development. *Mar Geol* 213(1–4):201–234
- Haflidason H, Lien R, Sejrup HP, Forsberg CF, Bryn P (2005) The dating and morphometry of the Storegga Slide. *Mar Pet Geol* 22(1–2):123–136
- Harbitz CB (1992) Model simulations of tsunamis generated by the Storegga Slides. *Mar Geol* 105(1–4):1–21
- Harbitz CB, Løvholt F, Pedersen G, Masson DG (2006) Mechanisms of tsunami generation by submarine landslides: a short review. *Nor J Geol* 86(3):255–264
- Heezen B, Ewing M (1952) Turbidity currents and submarine slumps, and the 1929 Grand Banks earthquake. *Am J Sci* 250:849–873
- Hill J, Collins GS, Avdis A, Kramer SC, Piggott MD (2014) How does multiscale modelling and inclusion of realistic palaeobathymetry affect numerical simulation of the Storegga Slide tsunami? *Ocean Model* 83:11–25

- Hoppe G (1965) Submarine peat in the Shetland Islands. *Geogr Ann* 47A:195–203
- Indrelid S (1973) En mesolitisk boplass i Dysvikja på Fjærtøft. *Arkeo* 1:7–11
- Indrelid S (1974) C-14 datering av boplassen i Dysvikja på Fjærtøft. *Arkeo* 1:10
- Jankaew K, Atwater BF, Sawai Y, Choowong M, Charoentitirat T, Martin ME, Prendergast A (2008) Medieval forewarning of the 2004 Indian Ocean tsunami in Thailand. *Nature* 455(7217):1228–1231
- Jansen E, Befring S, Bugge T, Eidvin T, Holtedahl H, Sejrup HP (1987) Large submarine slides on the Norwegian continental margin: sediments, transport and timing. *Mar Geol* 78(1–2):77–107
- Kaland PE (1984) Holocene shore displacement and shorelines in Hordaland, Western Norway. *Boreas* 13:203–242
- Kvalstad TJ, Andresen L, Forsberg CF, Berg K, Bryn P, Wangen M (2005) The Storegga slide: evaluation of triggering sources and slide mechanics. *Mar Pet Geol* 22(1–2):245–256
- Laberg JS, Vorren TO (2000) The Trænadjupet Slide, offshore Norway—morphology, evacuation and triggering mechanisms. *Mar Geol* 171(1–4):95–114
- Long D, Dawson AG, Smith DE (1989a) Tsunami risk in northwestern Europe: a Holocene example. *Terra Nova* 1(6):532–537
- Long D, Smith DE, Dawson AG (1989b) A Holocene tsunami deposit in eastern Scotland. *J Quat Sci* 4(1):61–66
- Long AJ, Barlow NLM, Dawson S, Hill J, Innes JB, Kelham C, Milne FD, Dawson A (2016) Lateglacial and Holocene relative sea-level changes and first evidence for the Storegga tsunami in Sutherland, Scotland. *J Quat Sci* 31(3):239–255
- Løvholt F, Harbitz CB, Haugen KB (2005) A parametric study of tsunamis generated by submarine slides in the Ormen Lange/Storegga area off western Norway. *Mar Pet Geol* 22(1–2):219–231
- Løvholt F, Bondevik S, Laberg JS, Kim J, Boylan N (2017) Some giant submarine landslides do not produce large tsunamis. *Geophys Res Lett* 44(16):8463–8472
- Micallef A, Masson DG, Berndt C, Stow DAV (2007) Morphology and mechanics of submarine spreading: a case study from the Storegga Slide. *J Geophys Res Earth Surf* 112 F03023. <https://doi.org/10.1029/2006JF000739>
- Rasmussen H, Bondevik S, Corner GD (2018) Holocene relative sea level history and Storegga tsunami run-up in Lyngen, northern Norway. *J Quat Sci* 33(4):393–408
- Rødal E (1996) Bønder og fiskere i Bud prestegjeld ca. 1600–1800, Romsdal Sogelag Årsskrift 1996. Romsdal Sogelag, Molde, pp 155–188
- Romundset A, Bondevik S (2011) Propagation of the Storegga tsunami into ice-free lakes along the southern shores of the Barents Sea. *J Quat Sci* 26(5):457–462
- Rydgren K, Bondevik S (2015) Moss growth patterns and timing of human exposure to a Mesolithic tsunami in the North Atlantic. *Geology* 43(2):111–114
- Shennan I, Horton B, Innes J, Gehrels R, Lloyd J, McArthur J, Rutherford M (2000) Late Quaternary sea-level changes, crustal movements and coastal evolution in Northumberland, UK. *J Quat Sci* 15(3):215–237
- Shi S (1995) Observational and theoretical aspects of tsunami sedimentation. Coventry University, Coventry
- Sissons JB, Smith DE (1965) Peat bogs in a post-glacial sea and a buried raised beach in the western part of the Carse of Stirling. *Scott J Geol* 1:247–255
- Smith DE, Cullingford RA, Haggart BA (1985) A major coastal flood during the Holocene in eastern Scotland. *Eiszeitalter und Gegenwart* 35:109–118
- Smith DE, Firth CR, Turbayne SC, Brooks CL (1992) Holocene relative sea-level changes and shoreline displacement in the Dornoch Firth area, Scotland. *Proc Geol Assoc* 103(Part 3):237–257
- Smith DE, Shi S, Cullingford RA, Dawson AG, Dawson S, Firth CR, Foster IDL, Fretwell PT, Haggart BA, Holloway LK, Long D (2004) The Holocene Storegga Slide tsunami in the United Kingdom. *Quat Sci Rev* 23(23–24):2291–2321
- Smith DE, Foster IDL, Long D, Shi S (2007) Reconstructing the pattern and depth of flow onshore in a palaeotsunami from associated deposits. *Sediment Geol* 200(3–4):362–371
- Solheim A, Berg K, Forsberg CF, Bryn P (2005a) The Storegga Slide complex: repetitive large scale sliding with similar cause and development. *Mar Pet Geol* 22(1–2):97–107
- Solheim A, Bryn P, Sejrup HP, Mienert J, Berg K (2005b) Ormen Lange—an integrated study for the safe development of a deep-water gas field within the Storegga Slide Complex, NE Atlantic continental margin; executive summary. *Mar Pet Geol* 22(1–2):1–9
- Svendsen J (1985) Strandforskivning på Sunnmøre. Bio- og litostratigrafiske undersøkelser på Gurskøy, Leinøy og Bergsøy. Bergen, Bergen, 142pp
- Svendsen JI, Mangerud J (1987) Late Weichselian and holocene sea-level history for a cross-section of western Norway. *J Quat Sci* 2(2):113–132
- Svendsen JI, Mangerud J (1990) Sea-level changes and pollen stratigraphy on the outer coast of Sunnmøre, western Norway. *Nor J Geol* 70:111–134
- Tjemsland A (1983) Vegetasjonshistoriske og paleolimnologiske undersøkelser av Rekkingedalstjørna og Sengsvatnet, Fedje, Hordaland. Bergen, Bergen, 213pp
- Vasskog K, Waldmann N, Bondevik S, Nesje A, Chapron E, Ariztegui D (2013) Evidence for Storegga tsunami run-up at the head of Nordfjord, western Norway. *J Quat Sci* 28(4):391–402
- Waddington C, Wicks K (2017) Resilience or wipe out? Evaluating the convergent impacts of the 8.2ka event and Storegga tsunami on the Mesolithic of northeast Britain. *J Archaeol Sci Rep* 14:692–714
- Wagner B, Bennike O, Klug M, Cremer H (2007) First indication of Storegga tsunami deposits from East Greenland. *J Quat Sci* 22(4):321–325
- Weninger B, Schulting R, Bradtmoller M, Clare L, Collard M, Edinborough K, Hilpert J, Joris O, Niekus M, Rohling EJ, Wagner B (2008) The catastrophic final flooding of Doggerland by the Storegga Slide tsunami. *Documenta Praehistorica* 35:1–24

Tsunamis Effects in Man-Made Environment

Harry Yeh, Andre Barbosa and

Benjamin H. Mason

School of Civil and Construction Engineering,
Oregon State University, Corvallis, OR, USA

Article Outline

Glossary

Definition

Introduction

Observations of Tsunami Damage

Tsunami Characteristics

Tsunami Generation, Propagation, and Run-Up
Modeling

Predicting Tsunami Forces

Remarks

Future Directions

Bibliography

Glossary

Bore A broken wave with an infinite wavelength that propagates into shallower quiescent water.

Buoyant force A net hydrostatic force on an object in the vertical direction.

Coastal armoring A human-made structure built to protect shorelines, e.g., seawalls, coastal dikes, and breakwaters.

Debris impact force Forces on an object caused by the impact of waterborne missiles.

Distant tsunami A tsunami created on a far away source; distant tsunami travel times can be on the order of hours.

Drag force Forces acting on a submerged object in the direction of a steady fluid flow.

Form drag A component of the drag force due to pressure difference between the front and back surfaces of the object.

Hydrodynamic force A force acting on a partially submerged structure by a steady free-surface flow around it.

Hydrostatic force Fluid forces under the uniform-flow condition with no vertical acceleration.

Local tsunami A tsunami created on a nearby source.

Momentum flux The steady portion of the net inertial forces.

Shallow-water-wave theory Water-wave theory with the assumptions of infinitesimal water depth relative to the wavelength for the irrotational fluid motion; alternatively, with the assumptions of hydrostatic pressure field and the inviscid-fluid flow with uniform horizontal velocity profile over the depth.

Shoaling effect Describes the increase in wave height as waves propagate into shallower water; it is a consequence of the conservation of wave energy flux.

Subduction zone A tectonic plate margin characterized by one plate going underneath (or subducting) of another tectonic plate; a plate dislocation in the subduction zone creates large tsunamis.

Surge force (impulsive force) Initial water impact forces on an object caused by the leading edge of the wave's upsurge motion on land.

Tsunami A water wave created by an impulsive disturbance in the ocean; the Japanese word "tsunami" translates directly to English as "harbor wave."

Vertical evacuation structure A human-made structure that provides a high-elevation safe haven from floods.

Definition

Tsunamis are water waves created by an impulsive disturbance in the ocean. A majority of

tsunamis are generated by a coseismic ocean floor displacement (i.e., earthquake-induced seafloor rupture). Less common generation mechanisms are subaerial or submarine landslides, volcanic activity, or, even more rarely, meteorite impact. To generate a tsunami, the earthquake must have a large magnitude (likewise, landsliding and volcanic activity must be significant to generate a tsunami). Because tsunami generation occurs when large spans of the seafloor displace (i.e., seafloor displacement of hundreds of kilometers), tsunamis' wavelengths are longer than the ocean depth. Tsunami wave heights are small (typically less than 1 meter high) in deep water, but the waves increase in height as they advance onto the continental shelf and approach the shore. The increase in wave height as waves approach shallow water is referred to as the shoaling effect. Tsunamis are capable of penetrating into well-protected harbors in coves, bays, or even fjords, because of their very long wavelength. Together with the shoaling effect, the sudden appearance of water rise in harbors is the origin of the Japanese word 津波 or "tsunami." The direct translation of tsunami in English is "harbor wave." Tsunamis can have significant impacts on coastlines and can damage or destroy coastal structures. Critical coastal infrastructure, such as nuclear power plants, oil and liquefied natural gas (LNG) storage and refinery facilities, ports and harbors, airports, coastal bridges, and vertical evacuation shelters, can be significantly impacted by impinging tsunamis. Failure of critical coastal infrastructure leads to loss of life and economic downtime for coastal communities. Notably, in this entry, we will focus on earthquake-induced tsunamis, though the discussions are applicable to tsunamis generated by other mechanisms.

Introduction

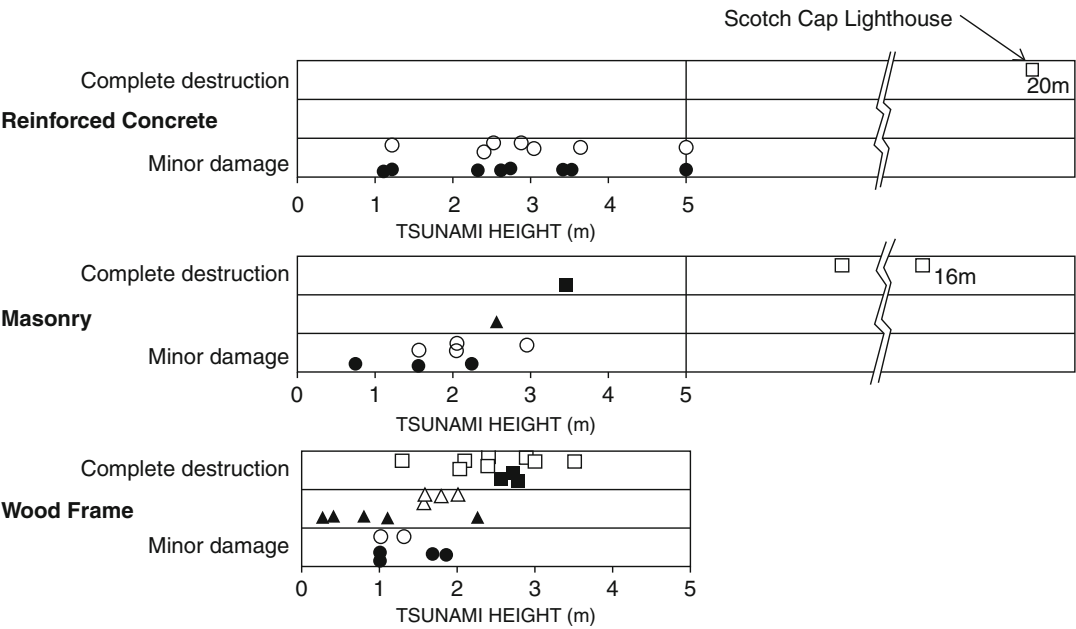
During fault rupture, seismic waves from the earthquake source propagate outward relatively quickly. An empirical seismological rule, which holds for many crustal earthquakes, states that the time between the leading arrival of a P-wave at a given site and the subsequent arrival of more

damaging S-wave (i.e., start of strong shaking) is approximately $D/8$, i.e., $t_s - t_p \approx D/8$, where t_s and t_p are the time of arrival of the S-wave and P-wave in seconds, respectively, and D is the source-to-site distance in kilometers (Lay and Wallace 1995). As an example, a source-to-site distance of 100 km, which is a characteristic source-to-coast distance for tsunami-generating earthquake sources, yields a $t_s - t_p$ of approximately 12 s. Accordingly, the largest lead-time warning that coastal areas can receive after earthquake rupture and before strong shaking starts is on the order of 10 s. This example highlights a key difference between earthquakes and tsunamis. Earthquake-induced tsunamis attack the shore tens of minutes or hours after the earthquake occurs, depending on whether the earthquake source is local or distant; thus, tsunami warning and evacuation, unlike earthquake warning and evacuation, are possible.

Currently, the primary tsunami-damage mitigation strategy is evacuation, because significant tsunamis are rare for a given locality and forewarning is possible. Accordingly, most tsunami-damage mitigation efforts have focused on effective warning systems, inundation mapping, and tsunami awareness education to promote effective evacuation strategies (National Tsunami Hazard Mitigation Program 2014).

Another tsunami-damage mitigation strategy, which started gaining popularity after the 2004 Indian Ocean Tsunami, is building tsunami-resistant coastal structures. Shuto (1994) compiled historical records of tsunami-induced building damage and found that reinforced concrete (RC) buildings withstood a majority of tsunami attacks. Shuto only reported one exception to the aforementioned observation: the total destruction of Scotch Cap Lighthouse by the 1946 Aleutian Tsunami (see Fig. 1). Such historical data were the basis for our dated understanding that "well-engineered" RC structures were "safe" from tsunami damage.

The 2011 Great East Japan Tsunami (hereinafter referred to as the "2011 Tsunami") attacked a developed, well-prepared country and led to widespread, unexpected damage; therefore, it is changing our understanding of tsunami



Tsunamis Effects in Man-Made Environment, Fig. 1 Degrees of building damage versus tsunami run-up height. This chart was compiled after the 1993 Okushiri

Tsunami by Shuto (1994). The marks filled in *black* are the data from the 1993 Okushiri tsunami, and the hollow marks are the data from the previous tsunami events

effects on coastal structures. For instance, one of the striking and unprecedented effects of the 2011 Tsunami was the complete failure of well-engineered RC coastal buildings. Post-tsunami reconnaissance showed that many RC buildings failed by overturning and sliding (some slid many meters away from their original location), and some of the toppled RC buildings had pile foundations. Excessive tsunami-induced damage was also observed in port and harbor facilities as well as coastal protective structures (e.g., seawalls, coastal dikes). Finally, the nuclear accident at the Fukushima Daiichi Nuclear Power Plant was tsunami-induced.

The 2011 Tsunami has clearly highlighted our lack of understanding of tsunami effects on coastal structures and, in particular, tsunami effects on critical structures, such as major coastal bridges, oil and LNG storage facilities, nuclear and fossil fuel power plants, military and civilian ports, and buildings used for vertical evacuation from tsunamis. Failure of critical structures causes enormous economic setbacks and triggers

additional hazardous situations that threaten human lives. The accelerating rate of construction of critical structures in coastal zones demands better understanding and development of a reliable methodology for design and assessment of tsunami-resistant structures. The 2011 Tsunami has also provided invaluable data and video footage that we can use to improve tsunami engineering practices worldwide.

There is an urgent need to develop methodologies, guidelines, and codes for tsunami-resilient buildings and infrastructures. This entry reports the present status of our understanding of tsunami effects on coastal structures and methodologies to estimate tsunami-induced forces on structures (noting the limitations, assumptions, and uncertainties of the methodologies). In addition, we discuss lessons learned from the past tsunami observations, tsunami characteristics and behaviors, classification of tsunami loadings on structures, and possible tsunami-induced failure mechanisms. We lean heavily on the 2011 Tsunami case history to frame this entry.

Observations of Tsunami Damage

Tsunami damage in coastal areas results primarily from the following causes: (1) flooding, (2) water forces, (3) impacts of waterborne debris, (4) scour and soil instability, (5) fire spread by floating materials, and (6) tsunami-induced wind.

Flooding from tsunamis is a major cause of damage to coastal structures. For instance, flooding from the 2011 Tsunami is thought to be the primary cause of the total power station black-out that cut off the cooling water supply and led to the nuclear meltdown at the Fukushima Daiichi Nuclear Power Plant. The operational nuclear reactors successfully shut down after the strong ground shaking ceased, and although the earthquake shaking damaged the electric grid, emergency generators successfully started to circulate the reactor coolant; therefore, the nuclear reactors were seismically robust (Disaster Countermeasures Office 2011). However, the reactors were vulnerable to the unexpectedly large tsunami flooding, which measured more than 14 m at the power plant site (Miller et al. 2011). The water damage from tsunami flooding disabled nearly all of the backup electric generators and the heat exchangers on site. As a result, the functional reactors were no longer cooled, which triggered the catastrophic nuclear core meltdown, in what is known as a loss of coolant accident (LOCA). The

Fukushima Daiichi Nuclear Power Plant LOCA is one of the most consequential infrastructure calamities caused by tsunami flooding in recent history. Importantly, tsunami flooding also causes uplift and transport of building structures as the buoyant force exceeds the weight of the buildings. Figure 2 shows a transported wood-frame house on Kunashir Island, Russia, during the 1994 Shikotan Tsunami. The house drifted 300 m onshore from its original location. Note that the house itself underwent minor structural damage.

Tsunami flows cause large water forces that can destroy and wash away wood-frame and masonry buildings as shown in the photo (Fig. 3) from the 1992 Nicaragua Tsunami. In Fig. 3, measured tsunami run-up heights range from 6.4 to 9.9 m (Abe et al. 1993). The debris from the tsunami-induced destruction of buildings is transported away from the original location. In contrast, debris caused by earthquake ground shaking remains at the original location. The difference in debris movement causes complications for post-tsunami reconnaissance efforts compared to post-earthquake reconnaissance efforts.

Tsunamis are also capable of transporting floatable objects, which can collide with other structures and cause damage. Figure 4 shows a large fishing boat (the 18th Kyotoku-maru, 60 m long, 330 t), which was carried 750 m from the port during the 2011 Tsunami. The boat destroyed

Tsunamis Effects in Man-Made Environment,

Fig. 2 In Yuzhno-Kurilsk, Kunashir Island, Russia, during the 1994 Shikotan Tsunami, this house drifted 300 m inshore along the riverbed (Photo by Harry Yeh)



Tsunamis Effects in Man-Made Environment,

Fig. 3 In El Transito, Nicaragua, during the 1992 Nicaragua Tsunami. This building and many surrounding buildings were washed away by tsunami, and tsunami debris was deposited further inland. The photo is looking toward the inland direction (Photo by Harry Yeh)



Tsunamis Effects in Man-Made Environment,
Fig. 4 At Kesennuma, Japan, during the 2011 Tsunami. This large fishing boat (the 330-t 18th Kyotoku-maru) was

transported 750 m from the port (Photo by Sankei Shinbun; map by Google Earth)

all the buildings in its path as it was transported by the tsunami (see the right panel of Fig. 4 for the relative location of the boat from the shoreline).

Tsunami inundation causes scour around buildings. Figure 5 shows a typical scour formation caused by the 2011 Tsunami at the corner of an apartment building in Yuriage, Japan. The building shown in Fig. 5 was approximately 800 m from the beach. Note that the scour hole must have formed in the location originally

covered by the asphalt pavement. The run-up height at this location was approximately 7.5 m (Mori et al. 2012).

Tsunamis are capable of rapidly spreading fires. The 1964 Great Alaska Earthquake and Tsunami caused a massive fire at the Texaco Petroleum storage tank facility in Seward, Alaska. The 1993 Okushiri Tsunami caused significant fires in the town of Aonae, Japan (see Fig. 6a). The Aonae fires were ignited by collapsing propane tanks,

Tsunamis Effects in Man-Made Environment,

Fig. 5 Scour caused in front of the apartment building in Yuriage, Japan, by the 2011 Tsunami (Photo by Harry Yeh)



and the tsunami spread the fire quickly. The 2011 Tsunami also caused fires in many locations. Figure 6b shows a fire one day after the tsunami near Kesennuma, Japan. Numerous floating debris pieces mixed with spilled oil are burning on the sea surface. Figure 6c shows a spilled oil sheet that filled the Ofunato Bay after the 2011 Tsunami.

Tsunamis' wave action can create significant wind, and the wind can damage coastal houses, although tsunami-induced wind damage is not common. Figure 7 shows the wind damage caused by the 1993 Okushiri Tsunami in Japan. A sudden wave deformation during tsunami run-up, together with the fast moving steep wave front, is thought to have generated winds strong enough to rip the house's siding. It is worth noting that this finding is based on interviews with the inhabitants of the house, who clearly stated that the damage did not originate during strong ground shaking. More details can be found in Shuto (1994).

Once tsunamis rip trees, destroy buildings, and pick up floatable objects (automobiles and boats), the resulting debris is transported and deposited. Tsunami debris can block critical transportation routes (roads, bridges, railroads, and ports and harbors), which causes significant delays of rescue missions and hampers efforts for fighting fires. Figure 8 shows the town of Rikuzentakata, Japan, before and immediately after the 2011 Tsunami as well as in August 2014. Debris originated from the destroyed houses and trees accumulated

at the outer edge of the town, which blocked transportation and delayed crucial rescue missions.

Tsunamis also undermine road pavement and lift pavement slabs (concrete or asphalt). Figure 9 shows a parking lot damaged by the tsunami in Yoriishohama, Japan, following the 2011 Tsunami. The observed parking lot damage could be caused by rapid fluctuations of tsunami inundation and drawdown.

Figure 10 shows the change in coastal geomorphology caused by the failure of a seawall near Murakami Beach in Japan following the 2011 Tsunami. Breaching a short section of the seawall created a clean, crescent-shaped bay formation with a diameter of approximately 60 m.

The foregoing tsunami-damage descriptions indicate that the damage patterns are quite distinct from other natural hazards. This is partly because a tsunami hazard is a relatively rare event, and tsunami loadings have unique behaviors and characteristics. Therefore, both natural and human-made systems are not typically prepared for tsunamis, even if the tsunami's forcing magnitude is small. Details of tsunamis' unique behaviors and characteristics are discussed next.

Tsunami Characteristics

Tsunamis have unique characteristics compared to other natural hazards. Primarily, tsunamis'



Tsunamis Effects in Man-Made Environment, Fig. 6 Fire damage caused by tsunamis: (a) Aonae, Japan, during the 1993 Okushiri Tsunami, note that a majority of the town was burned when propane gas canisters ignited and the fires were subsequently spread by the tsunami (Photo by Dennis J. Sigrist); (b) Ninohama, Japan,

approximately 18 h after the tsunami, the sea surface is burning with numerous debris pieces mixed with spilled oil; (c) oil spill caused by tsunamis in the port of Ofunato, Japan, approximately 18 h after the 2011 Tsunami (The photos in (b) and (c) were enhanced by Harry Yeh from the originals taken by Kenji Satake)

temporal and spatial characteristic scales are different from the temporal and spatial characteristic scales of other natural hazards. Table 1 shows typical characteristics of different types of natural hazards. Notably, the timescale of tsunami inundation is different from other hazards; hence, tsunami inundation behavior is distinct.

River floods are characterized by a gradual increase and decrease in water level. Although locally enhanced transient flows occur when riverbanks are overtopped or breached, the flow within the main river channel varies slowly in time. A storm surge creates the condition of relatively short-period waves (on the order of a 10s wave period) riding on the gradual increase and decrease of the mean water level in the coastal zone. The maximum run-up height during a storm

surge (n.b., the maximum run-up height is the difference between the water elevation at the maximum inundation penetration and the sea level at the time of attack) is controlled by the gradually varying “surge component”; i.e., the short-period waves are dissipated and cause a minor effect on the maximum run-up penetration. As a result, the maximum storm surge inundation is usually distributed fairly uniformly along the coastal line, which is not the case for tsunamis. Tsunamis’ spatial and temporal scales are intermediate compared to other natural hazards (Table 1). The combination of a long-period wave occurring over a temporal scale of minutes and interacting with the likely complex bathymetry and topography leads to unusual behavior. Consequently, tsunami run-up heights often exhibit significant variability



Tsunamis Effects in Man-Made Environment, Fig. 7 In Aonae, Japan, during the 1993 Okushiri Tsunami. This house's top story was damaged with no tsunami water contact (tsunami inundation only reached the half height of the first floor) and no debris impact; therefore, a possible cause of the damage is strong wind induced by the tsunami run-up (Photo by Nobuo Shuto)

along a relatively short and narrow coastal zone. Figure 11 shows significant variation in the run-up heights measured along 4 km of the northwest coastline of Okushiri Island, Japan, during the 1993 Okushiri Tsunami.

Every year, tens of hurricanes/typhoons are spawned, and some make landfall, which may lead to damage in coastal areas. River floods occur relatively frequently. According to the US Geological Survey, potentially damaging earthquakes, with magnitudes larger than 6.0, occur approximately 150 times per year worldwide (USGS 2014). In contrast, significantly damaging tsunamis are rare. Consequently, both nature and humans are not prepared for infrequent tsunamis and their disastrous effects. For example, the intruding tsunami currents of the 1960 Chilean Tsunami resulted in erosion of more than 10 m at the port entrance of Kesennuma in Japan (Takahashi et al. 1992). Before the 1960 Chilean Tsunami, the Kesennuma Port had experienced many large storms and typhoons, but none of the previous storms or typhoons had caused nearly as

much erosion as the tsunami. Based on the aerial photographs, the measured maximum flow speed of the 1960 Chilean Tsunami was less than 3 m/s (Takahashi et al. 1992), which is comparable to the maximum flow speed of a river flood. However, the Kesennuma Port entrance is not a river, and the sea bottom is characterized by soft sediments, which had not experienced unusual tsunami flows.

Tsunami-risk areas are limited to narrow regions along the shoreline. Within the tsunami-risk areas, damage and losses are not uniform; i.e., the nearer the shoreline, the higher the tsunami intensity. The same nearshore observations apply to hurricane/typhoon hazards. However, there are distinct differences between tsunamis and storm surges created by hurricanes. Storm surges created by hurricanes are limited to the area directly affected by the storms; in most cases, the affected coastline is a few hundred kilometers in length. In contrast, tsunamis can affect the coastlines of entire oceans. Note that storm-generated waves are dispersive, because they have wavelengths shorter than the ocean depth. Because the speed of wave propagation depends on the wavelength (or frequency), a waveform that consists of many frequency components, such as a storm wave, spreads its energy in the propagation direction. Storm-generated waves thereby continually reduce in amplitude during propagation, which is why they are considered dispersive. On the other hand, tsunamis have very long wave periods; therefore, tsunamis can propagate long distances without dispersing strength and can cause significant damage far away from the source. Accordingly, coastal areas having no local tsunami trigger (e.g., earthquakes, submarine landslides) can be attacked by "distant tsunamis." In the past when modern communication devices did not exist, distant tsunamis were devastating, because no forewarning was available to the residents. For example, the 1960 Chilean Tsunami attacked Japan, which is 17,000 km away, and killed 229 people within the Tohoku region (Takahashi and Hatori 1961). Tohoku is the region of Japan also the most affected by the 2011 Tsunami. A more recent case is the 2004 Indian Ocean Tsunami, which caused a large number of



Tsunamis Effects in Man-Made Environment, Fig. 8 Past (*top left*), immediately after (*top right*), and present (*bottom*; August 2014) scenes of Rikuzentakata, which was affected by the 2011 Tsunami. Debris causes

serious blockages, which limits emergency responder access to the affected areas (All images taken from Google Earth)

Tsunamis Effects in Man-Made Environment, Fig. 9 Ripped pavement surface caused by the 2011 Tsunami. This area was a parking lot for the fishing port of Yoriisohama, Japan. The damage was presumably caused by a combination of rapid fluctuations in pore-water pressures, skin friction, and lift forces (Photo by Harry Yeh)



Tsunamis Effects in Man-Made Environment,

Fig. 10 A miniature crescent bay formation in Murakami Beach, Fukushima, Japan, due to a breached seawall by the 2011 Tsunami (Photo by Harry Yeh)

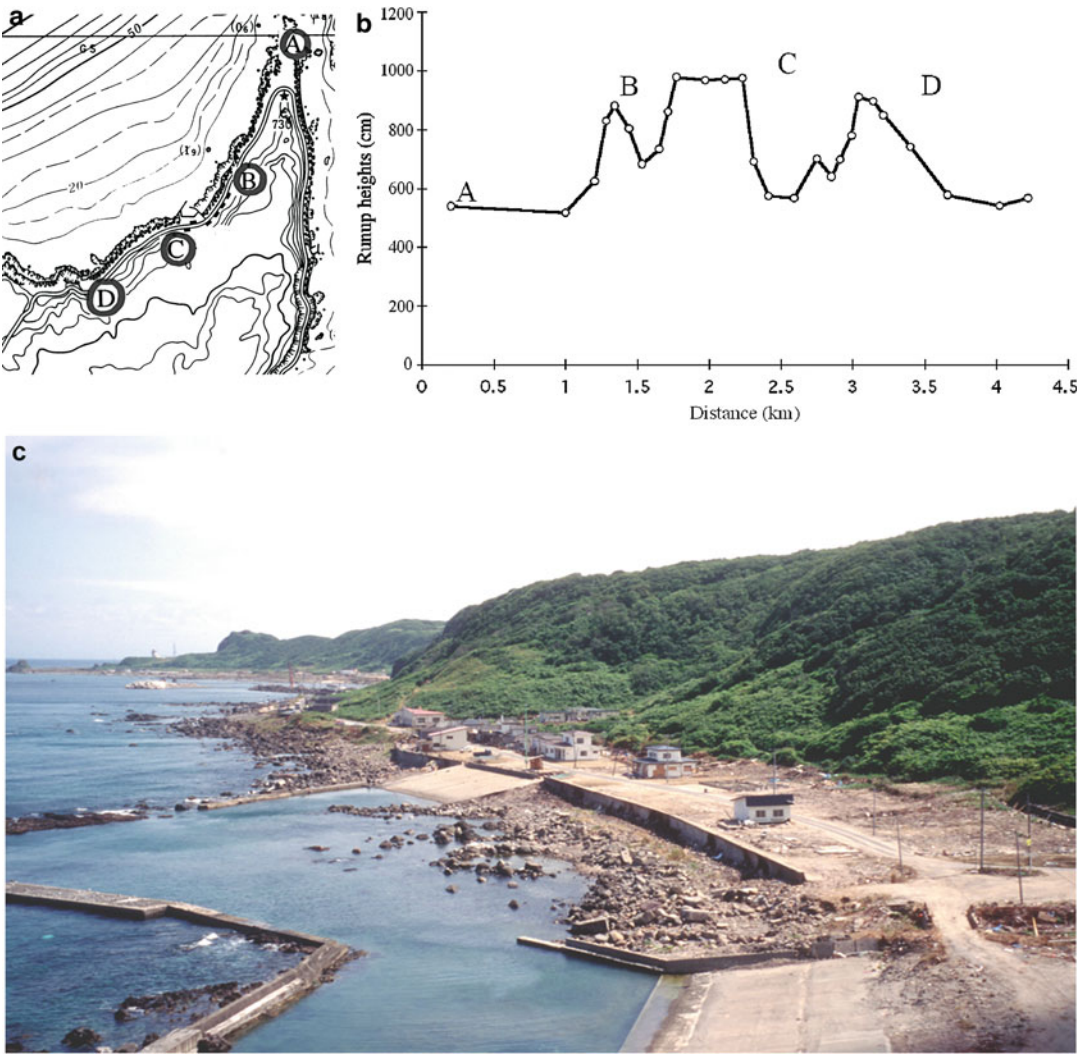


Tsunamis Effects in Man-Made Environment, Table 1 Timescale and length scale characteristics of several common natural hazards

Hazards	Timescale	Length scale	Typical pressure head	Lead time
River flood	Days	100 km	3 m	Hours to days
Hurricane/storm surge	Hours	1,000 km	3 m	Days
Storm-generated waves	Seconds	$\ll 1$ km	10 m	Days
Tsunami	Minutes	1 km ~ 500 km	10 m	Minutes to hours
Earthquake	Seconds	N/A	N/A	Seconds

fatalities in distant countries. The number of confirmed deaths caused by the 2004 Indian Ocean Tsunami in Sri Lanka, India, and Thailand is 31,147, 10,749, and 5,395, respectively (Wachtendorf et al. 2006). The 2004 Indian Ocean Tsunami was a devastating natural disaster that affected many lives; however, the tsunami motivated many advances in tsunami monitoring. Consider, for example, the Deep-ocean Assessment and Reporting of Tsunamis (DART) system developed by the National Oceanic and Atmospheric Administration (NOAA). The DART system has improved communication of tsunami hazards, has created a global tsunami warning system, and has increased public awareness of tsunamis. The DART system, among other tsunami warning technologies, has enabled us to issue timely warnings, which promotes effective evacuation strategies for distant tsunami events and saves human lives.

Tsunamis often cause significant casualties, because populations are poorly prepared for such a rare and unique natural disaster. As an example, the 2004 Indian Ocean Tsunami destroyed buildings and killed healthy adult humans in Devanampattinam, India, even though the tsunami flow depth was approximately 1.5 m (Yeh et al. 2007). In contrast, storm-generated waves of equivalent height would not have caused the same damage, because the storm waves would have a period of about 10 s or so. Even if a human were swept away by a storm wave, she or he would likely survive after one wave period in the inundation zone. On the other hand, a tsunami's very long wave period causes very long and sustained penetration (a tsunami's period can vary from 10 to 120 minutes or longer). Unfortunately, a human body, once swept away by a tsunami, may not resurface during the sustained tsunami action.



Tsunamis Effects in Man-Made Environment, Fig. 11 Measured run-up heights of the 1993 Okushiri Tsunami along Inaho Coast, Japan, which demonstrates

that the run-up height varies significantly within the neighboring region (Photo by Harry Yeh)

Distant-source tsunamis can induce currents with significant flow velocities. Okal et al. (2006) reported that, during the 2004 Indian Ocean Tsunami, the strong currents in the Port of Salalah, Oman (approximately 4,800 km away from the tsunami source in Indonesia), broke all of the mooring lines on a 285-m-long ship. After becoming unmoored, the ship dangerously drifted around the harbor before beaching on a nearby sandbar. More recently, strong currents induced by the 2011 Tsunami were measured in Crescent

City Harbor, California, and Pillar Point Harbor, California (Lynett et al. 2012).

Tsunamis react to smaller local features during inundation. Tsunami impacts on structures are substantially affected by the surroundings. Figure 12 shows damage in the town of Onagawa, Japan, immediately following the 2011 Tsunami. The damage pattern shown in Fig. 12 suggests that the sturdy waterfront buildings (a pair of brown-colored buildings in the photo) must have acted as a barrier for smaller buildings behind them

Tsunamis Effects in Man-Made Environment,

Fig. 12 Tsunami destruction pattern in Onagawa, Japan, due to the 2011 Tsunami. A pair of sturdy large buildings at the waterfront (highlighted by the arrows) acted like barriers for some of the weaker buildings behind. (This photo was enhanced by Harry Yeh from the original taken by Kenji Satake.) The gap between the buildings created a jet flow that completely destroyed the narrow strip within the jet trajectory and caused a deep scour hole



(i.e., the smaller buildings were in the “tsunami shadow” of the larger buildings). Video footage taken in Onagawa during the tsunami shows a strong water jet formation in the gap between the two large buildings (many of the videos are available on YouTube). In other words, the gap of the sturdy large buildings formed a “nozzle,” which increased the flow velocity substantially. As a result, more damage was observed to buildings within the jet formation. The presence of the sturdy buildings substantially altered the tsunami flow velocities, which in turn affected the surroundings. In addition, the scouring effect is increased as the tsunami flow velocity increases, which led to more damage to coastal structures. This example from Onagawa highlights the difficulty of predicting tsunami effects over the scale of cities.

One of the lessons learned from the 2011 Tsunami is that tsunamis can cause severe damage to coastal protection structures. Coastal protection structures are traditionally designed to protect coastal areas from storm surges (note that there are, however, a limited number of coastal protection structures in Japan designed for historical tsunamis). When tsunamis exceed the design height, the coastal protection structures can no longer provide total protection, although they

can still reduce the damaging effects of the tsunami. Figure 13 shows damage to a coastal dike in Isobe, Japan, caused by the 2011 Tsunami. The tsunami run-up height was 16.25 m (Mori and Takahashi 2012). A series of detached breakwaters was placed approximately 100 m offshore of the dike to provide extra protection. The pre-tsunami crown elevation of the seawall was 6 m relative to the Tokyo Pell datum; accordingly, the tsunami easily overtopped the seawall (n.b., Tokyo Pell, or TP, is the Japanese sea surface datum, and tsunami run-up heights are reported relative to the sea level elevation at the time of the tsunami attack, which was approximately 40 cm higher than TP when the 2011 Tsunami attacked). The gently sloped seaward face of the coastal dike was left intact after the tsunami, but significant damage occurred along the crown and along the landward slope. A large amount of water overtopped the seawall, which caused a huge scour strip to form, as shown in Fig. 13. The scour strip behind the coastal dike was a thick pine tree forest prior to the 2011 Tsunami. According to Nakao et al. (2012), the observed damage pattern can be explained with the effect of centrifugal forces induced by the curved flows near the crown of the dike (convex flow) and the toe of the dike (concave flow). Near the crown, the induced

Tsunamis Effects in Man-Made Environment,

Fig. 13 Coastal dike failure in Isobe, Japan, due to the 2011 Tsunami. Note that there is minor damage on the front face, the capstones of the crown were wiped away, and a severe scour hole formed on the lee side of the dike where there used to be a thick pine tree forest (Photo by Harry Yeh)



negative force is significant enough to pull out the concrete caps. In contrast, the significant positive force and fast moving flows at the toe must have created the significant scour strip.

Tsunami Generation, Propagation, and Run-Up Modeling

Accurate quantitative predictions of tsunami effects on coastal structures are a formidable topic. Regardless, to cope with tsunami hazards, it is necessary to estimate the effects with our best knowledge and understanding. To predict tsunami effects, tsunami inundation depth $h(\vec{x}, t)$ and the tsunami inundation velocity $\vec{u}(\vec{x}, t)$ must be estimated, where $\vec{x}$ is the position vector of the location of interest and t is the time. The time histories of $h(\vec{x}, t)$ and $\vec{u}(\vec{x}, t)$ are needed to properly estimate wave forces on structures as well as tsunami-induced soil instability and the buoyancy force for partially buried structures. In the present practice, numerical models are used to obtain estimates of $h(\vec{x}, t)$ and $\vec{u}(\vec{x}, t)$. Numerical simulations of tsunamis involve several stages: (1) tsunami generation, which defines the initial condition, (2) tsunami propagation, and (3) run-up to the land. With given seismic parameters (i.e., seismic moment, locations of the epicenter and the hypocenter, and the seismological parameters such as the strike, dip, and rake angles), the seafloor displacement can be computed with

linear elastic dislocation theory (e.g., Mansinha and Smylie 1971; Okada 1985). The prediction of seafloor deformation involves great uncertainty, because the seismic fault rupture process is inhomogeneous, and thus, the seismic parameters cannot be known with adequate accuracy.

Numerical simulations of tsunami propagation and run-up require accurate bathymetric and topographic data. Typical tsunami wavelengths in deep water are on the order of several tens to hundreds of kilometers; therefore, tsunami-induced flow can reach the seafloor, even in a 4,000-m-deep abyssal plain. Consequently, tsunami propagation is strongly affected by bottom bathymetry. Ocean bathymetry has not been completely measured, especially in the southern hemisphere; hence, incomplete bathymetry is another significant source of uncertainty for predictions of tsunami effects. Note that the same bathymetry-related concerns are usually not present for wind-generated waves, which are typically less than 500 m long, because waves with a wavelength less than twice the seafloor depth are minimally affected by the presence of the ocean bottom. Accordingly, numerical simulations of tsunami propagation and run-up require integrating bathymetry data over the entire ocean basin and dry coastal topography. The need for such a complete understanding of bathymetry and topography adds difficulty to the numerical simulations of tsunamis, because we need reliable bathymetric and topographic data for 10,000-m-deep ocean

trenches, 4,000-m-deep abyssal plains, 200-m-deep continental shelves, and complex coastal formations above sea level.

Although tsunami waves contain a broad spectrum of wave periods at the source, most of the energy is retained in the long-wave components as the tsunami propagates to the shore, and shorter-length (higher-frequency) waves disperse quickly, as discussed earlier. Shorter-length gravity-driven waves propagate more slowly than the longer-length gravity-driven waves. Given the dispersion of shorter-length waves, tsunami propagation is often computed based on the shallow-water-wave theory. The theory comprises the conservation of fluid volume and the conservation of depth-averaged linear momentum with the assumptions of a hydrostatic pressure field and uniform horizontal velocities over depth. Typical formulations of the theory can be expressed respectively as

$$\begin{aligned} \frac{\partial h}{\partial t} + \nabla \cdot (\vec{h}\vec{u}) &= 0 \\ \left(\frac{\partial}{\partial t} + \vec{u} \cdot \nabla \right) \vec{u} + g\nabla h + \gamma \frac{g|\vec{u}|\vec{u}}{h^{4/3}} &= g\nabla h_0, \end{aligned} \quad (1)$$

where h_0 is the water depth from the referenced datum (e.g., the quiescent water level), g is the gravitational acceleration, and γ is the friction coefficient. The resulting model is nondispersive in frequency; therefore, the propagation of wave energy (e.g., the group celerity) is independent of wave number (or wavelength). The use of shallow-water-wave theory to model tsunami wave propagation can be justified, because tsunamis from coseismic sources have very long wavelengths in comparison with the ocean depth. Wave nonlinearity is not prominent for tsunamis propagating in deep oceans, because the tsunami amplitude in deep water is very small (at most a meter or so) in comparison with the depth. Therefore, the linear shallow-water-wave theory works adequately for estimating the propagation of tsunami waves in deep oceans (Yeh et al. 1996).

When the tsunami wave intrudes onto the continental shelf, the wave amplitude increases due to the shoaling effect, and as a result, wave nonlinearity effects become important. The wave nonlinearity effect can be conceptualized with an energy argument. In deep water, a tsunami's kinetic energy is uniformly distributed throughout the deepwater column. As the tsunami propagates into shallower water, the deepwater column is "squeezed" to a shallow-water column, and some portion of the tsunami's kinetic energy (fluid velocity) is converted to potential energy (wave height). As the tsunami reaches the continental shelf, wave dispersion could become important depending on the wavelength of the incoming tsunami and the width of the continental shelf. When the continental shelf is sufficiently wide in comparison with the tsunami wavelength, a single pulse of the incoming tsunami can be transformed to a series of shorter waves due to the dispersion effect. However, when the continental shelf is narrow relative to the incident tsunami wavelength, there is no sufficient time for dispersion to occur, and thus, the tsunami reaches the shore with minor dispersion effects. When tsunami run-up occurs on the beach and further inland, friction effects and turbulence (associated with wave breaking) become important, and the run-up motion becomes intrinsically nonlinear. The run-up model also needs to consider natural and human-made shoreline configurations, e.g., buildings, trees, mounds, and roads.

The foregoing description of numerical simulation of tsunami generation, propagation, and run-up evidently demonstrates that the problem is complex and multi-scale. It is a complex problem because it involves multiphase (water, air, solid) interactions in a three-dimensional real-world domain where some fundamentals (e.g., turbulence) remain unsolved. It is a multi-scale problem because the length scale of tsunamis in the ocean is on the order of hundreds of kilometers, and the effects of inundation phenomena must be described at scales of a few meters or less. Hence, at present, even the best tsunami modeling involves substantial uncertainty.

Predicting Tsunami Forces

Fundamentally, fluid forces, $\vec{F}$, exerted on a non-compliant body are computed by

$$\vec{F} = \int_{\partial S} \tau_{ij} n_j ds, \quad (2)$$

where τ_{ij} is the stress tensor, n_j is the unit vector representing the direction of the surface ds , and ∂S is the net surface area of interest. The integrand in Eq. 2 can be split into two components, namely, the component normal to the surface ($i = j$), which is called the pressure, and the vector component tangent to the surface ($i \neq j$), which is called the shear stress. The two components are the basis of fluid loading on a stationary rigid body.

For engineering estimates, tsunami forces are classified based on the typical flow conditions; for example, FEMA P646 (2012) classifies the tsunami forces as (1) hydrodynamic forces under steady ambient flows, (2) hydrostatic and buoyant forces under quiescent flood states, (3) highly transient surge forces at the impact of the leading edge of the arriving water mass (alternately called “impulsive forces”), (4) debris impact forces resulted from waterborne missiles, and (5) debris damming forces. The classification is made for convenience for engineering applications, but the actual forces are represented by Eq. 2.

The term *hydrodynamic force* is often used in the tsunami literature (e.g., Dames and Moore 1980; FEMA P646 2012; Yeh 2006, 2007). The hydrodynamic force is similar to, but not exactly the same as, the *drag force* used in the field of fluid mechanics. The term drag force in fluid mechanics is used to indicate the force acting on a body fully submerged in a “steady” flow; hence, no air–water free surface is involved. On the other hand, the hydrodynamic force used in the tsunami literature indicates the fluid force exerted on a partially submerged body.

The drag force is the component of the total force parallel to the ambient flow direction. Correspondingly, the lift force is the component of the total force perpendicular to the ambient flow

velocity. If a body is fully submerged in a steady flow, then the drag force, F_D , is calculated by

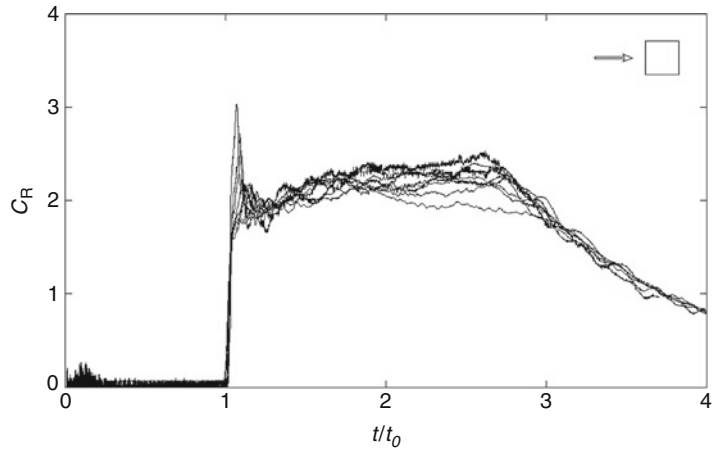
$$F_D = \frac{1}{2} C_D A \rho U^2, \quad (3)$$

where C_D is the coefficient of drag; A is the projectional area of the body on a plane normal to the direction of ambient steady flow, which has the velocity magnitude U ; and ρ is the fluid density. A portion of the drag force is due to skin friction (i.e., tangential viscous shear force along the body), and the remaining portion of drag force is due to the pressure distribution around the body, which is called *form drag*. Equation 3 was derived by dimensional analysis and validated by experiments (e.g., Batchelor 1967).

It is emphasized that the force given in Eq. 3 represents the drag force exerted on the submerged body as a whole and not the force exerted on only a portion of the submerged body. However, the form of Eq. 3 is often used for a partially submerged body, as noted above. Notably, the theoretical justification for the use of Eq. 3 for gravity-controlled free-surface flows is not rigorous, because the dimensional analysis used to derive Eq. 3 does not consider the gravity effect. Arnason et al. (2009) recognized this potential shortcoming, and, instead of the drag coefficient, C_D , they introduced “the coefficient of resistance,” C_R , to distinguish the forces on a partially submerged cylinder. Arnason et al. (2009) found that the form of Eq. 3 yielded good approximations to their laboratory experiments for a vertically erected square column (see Fig. 14), and the magnitude of C_R is comparable to C_D for a fully submerged square column; i.e., $C_R \approx 2$. It is anticipated that for a body that is “not-too-elongated” in the flow direction, the dominant force is the form drag, which is caused by the pressure difference (roughly equivalent to the water surface difference) in front and back of the body. In summary, Eq. 3, which is often used to calculate the hydrodynamic force, can be used to evaluate the net fluid force on a body in a steady (or quasi-steady) flow.

Tsunamis Effects in Man-Made Environment,

Fig. 14 Laboratory data of net horizontal forces of incident bores onto a vertical square column. The coefficient of resistance, C_R , represents the normalized force and is approximately equal to 2, except at the initial impingement. The multiple lines represent the cases with different incident bore strengths



The *hydrostatic force* is a common force discussed in fluid mechanics. A hydrostatic condition is achieved when the pressure force balances with the gravitational body force, namely,

$$\frac{\partial p}{\partial z} = -\rho g, \quad (4)$$

where p is the pressure and the coordinate z points upward (i.e., opposite to the gravitational direction). The hydrostatic condition exists in the absence of inertial and viscous forces. The hydrostatic force can be calculated by integrating the pressure field obtained in Eq. 4 over the surface of interest. Note that there is no net hydrostatic force in the horizontal directions on a whole body in quiescent water, if the integration of Eq. 2 is made around the entire body. The integration of the vertical pressure component produces a net upward force, which is called the buoyant force. The hydrostatic force is often calculated for the force acting on one side of a building component surface, e.g., walls and windows.

When a fluid is in motion around a body, a hydrostatic condition no longer exists around the body. Instead, vertical flow accelerations can form in the wake on the lee side of the body, and horseshoe vortices can form around the front face. It must be emphasized that when the fluid is in motion around a body, the force acting on the body is still a pressure force, but it should not be calculated using Eq. 4. Notably, when the vertical

flow accelerations at the body are small, which may occur during a gradual tsunami inundation process, the force on a portion of the body (e.g., windows, walls) can be approximated by assuming a hydrostatic condition, though the error involved in the approximation must be recognized.

The *buoyant force* is created by the net upward pressure acting on a partially or totally submerged body under a hydrostatic condition. The vertical buoyant force is computed by Eq. 2 using the hydrostatic force definition given in Eq. 4. The buoyant force is physically equivalent to the fluid weight displaced by a body, i.e., Archimedes' principle (e.g., Lamb 1960). To conceptualize the importance of the buoyant force for tsunami problems, consider a body that is partially buried in the ground, such as a building with a basement. To establish the upward pressure required to develop the buoyant force, the excess pore-water pressure in the soil must increase during tsunami inundation.

To consider the downward gradient of pore-water pressure, the excess pore-water pressure dissipation must be considered. Biot's (1941) poroelasticity equations can be used to develop this solution; however, Biot's equations are rarely used for this purpose, because they require physical parameters that are hard to measure or estimate. Instead, a simpler governing equation, Terzaghi's (1943) one-dimensional consolidation equation, is often used to predict excess pore-

water pressure dissipation. Note that Terzaghi's equation is a simplified formulation from Biot's equations. Terzaghi's equation is in the form of a diffusion equation with the coefficient, c_v , which is often referred to as the coefficient of consolidation. Terzaghi's equation is given by

$$\frac{\partial p_e}{\partial t} = c_v \frac{\partial^2 p_e}{\partial z^2}, \quad (5)$$

where p_e is the excess pore pressure, i.e., $p_e(t) = p(t) + \rho g z$. The coefficient of consolidation, c_v , can be expressed as (e.g., Holtz and Kovacs 1981)

$$c_v = \frac{k}{\rho g} \frac{1 + e_0}{a_v}, \quad (6)$$

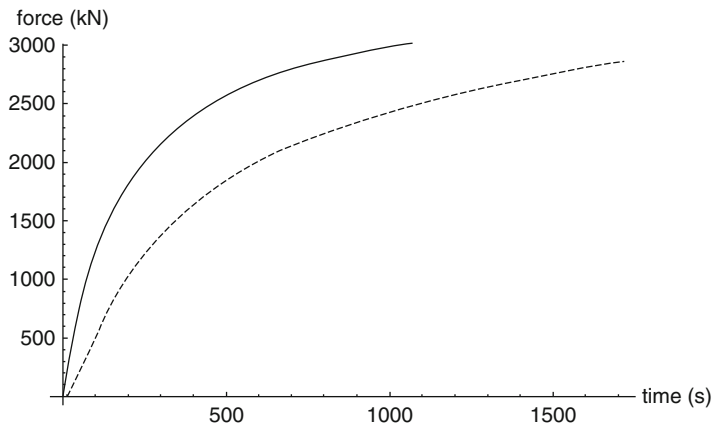
where k is the hydraulic conductivity in Darcy's law, e_0 is the void ratio when no external stress is applied, and a_v is the coefficient of compressibility, i.e., $a_v = -de/d\sigma'$, where e is the void ratio and σ' is the effective stress. The effective stress exerted on soil is defined as $\sigma' = \sigma - p$, i.e., the total stress, σ , minus the pore-water pressure, p . In hydrodynamics, it is more common to use the porosity, ϕ , instead of void ratio [$\phi = e/(e + 1)$] and the bulk modulus of elasticity, K , instead of the coefficient of compressibility ($K \approx 1/a_v$). Evidently, the values of k and a_v change during tsunami inundation, which indicates that c_v also changes, and consequently, Eq. 5 is nonlinear. For a natural soil deposit, the values of c_v are spatially variable. Despite these complications, a

temporally and spatially constant value of c_v is usually assumed for a given soil medium when using Eq. 5. Values of c_v for sandy soils, which are typically found in coastal areas, are seldom experimentally measured, because excess pore-water pressure usually dissipates quickly in sandy soils. Zen and Yamazaki (1990, 1991) experimentally measured c_v for different sands and found the range to be approximately 10 to 10,000 cm²/s. Clearly, Eq. 5 with its single diffusion coefficient has limitations for the prediction of excess pore-water pressure dissipation in sandy soils. Nonetheless, with a well-determined value of constant c_v , Eq. 5 could yield good approximations of pore-water pressure dissipation, at least qualitatively.

Note that the pore-water pressure generation and dissipation processes are not instantaneous; i.e., excess pore-water pressures require a finite time to establish during inundation. Consequently, the effect of the buoyant force on an embedded body must depend on (a) the duration and the depth of tsunami inundation and (b) the depth of the bottom surface of the embedded body. Figure 15 shows the time variation of the upward buoyancy force under a vertical cylindrical column of radius 5 m computed by Eq. 5 with a constant value of c_v . Figure 15 was created with the assumption of an instantaneous inundation load at $t = 0$ s. The results for the different embedment depths show that the buoyant force establishes more quickly for shallower embedment depths, which suggests that building with basements should be more robust to tsunami loading

Tsunamis Effects in Man-Made Environment,

Fig. 15 Buoyant force exerted under a cylindrical-shaped object with the radius 5 m under an instantaneous inundation of 5 m at $t = 0$. Solid line, the bottom face buried in 1.0 m (no basement); dashed line, the bottom face buried in 3.0 m (with basement). The soil has the consolidation coefficient $c_v = 700$ cm²/s



than buildings without basements, because large buoyant forces tend to destabilize buildings.

The term *surge force* was first used by Cross (1967). The surge force is referred to as the *impulsive force* in FEMA P646 (2012). It is caused by the impingement of a leading edge of surging water onto a body. The surging water on a tsunami's leading edge advances on a "dry" bed (i.e., partially saturated), which leads to a highly transient, multiphase (i.e., air, water, sediments) flow. The surge force acts only on the front side of the body, while the rear side of the body is still dry. Because the surge flow (and thus the surge force) is highly transient, it is important to consider ductility of the impacted body. The compliant response of the body reduces the net momentum exchange between the fluid motion and the solid body. Presently, there is no established and rational method available to predict the surge force with all of the aforementioned considerations. However, Cumberbatch (1960) presented a similarity solution of the initial impact (i.e., prior to when the gravitational acceleration starts affecting flow) of a fluid "wedge" onto a rigid wall by using the potential flow assumption. Cumberbatch found that the surge force was greater than the total momentum ($\rho U^2 A$) past the wall location by factors 2.4 and 1.6 for surging fronts with 45° and 22.2° semi-angle wedges, respectively. Following the work by Cumberbatch, Cross (1967) proposed to add the hydrostatic pressure force to the momentum flux to estimate the surge force per unit breadth:

$$F = C_F \rho h U^2 + \frac{1}{2} \rho g h^2, \quad (7)$$

where U is the speed of the incident fluid wedge, C_F is Cumberbatch's force coefficient ($C_F = 2.4$ and 1.6 for the 45° and 22.2° semi-angle wedges, respectively), and h is the water depth. Arithmetically, adding the hydrostatic force to Cumberbatch's analysis cannot be supported for nonlinear flow action. In addition, estimating the water depth for a highly transient, boundary-dominated flow at the leading edge of a surge is ambiguous. Other derivations from the original

Cross (1967) approach (e.g., Paczkowski et al. 2011) face similar problems; i.e., justifying the rationality of the approach and identifying unambiguous model parameters are both problematic.

Yeh (2007) found that surge forces are often smaller than the subsequent quasi-steady hydrodynamic forces. Yeh postulated that the relatively small water-surface slope of the leading edge of the surge when it advances on a dry bed is the primary reason for the surge forces being smaller than the quasi-steady hydrodynamic forces. Notably, Yeh's postulate is consistent with the work by Cumberbatch (1960); i.e., a smaller wedge semi-angle leads to a smaller value of Cumberbatch's force coefficient, C_F .

Yeh (2007) further studied this issue by examining bore propagation. A bore is a broken wave with an infinite wavelength that propagates into a uniform height region of quiescent water. Yeh commented that the front surface of a bore becomes steeper when the bore propagates in a water of finite depth. Note that the bore flow condition differs from surge flow condition, because surges advance on dry beds. Based on two independent laboratory studies by Ramsden (1993) and Arnason (2005), the upper limit of the initial impact force caused by a bore traveling in water with a finite depth is approximately 150 percent of the subsequent maximum hydrodynamic force in a quasi-steady flow.

Debris impact forces are more difficult to estimate than other tsunami-related forces. Several engineering attempts to estimate debris impact forces have been made in the past, and the attempts are summarized in Appendix D of FEMA P646 (2012). Unlike other forces, debris impact forces occur locally at the point of contact when the debris is smaller than the building. Debris impact forces are assumed to act at or near the water surface level when the debris strikes the structure. Some models for incorporating debris impact forces are based on the impulse-momentum principle, which states that the impulse of the resultant force acting for an infinitesimal time is equal to the change in linear momentum, i.e.,

$$I = \int_0^{\tau} F dt = d(mu); \quad \tau \rightarrow 0. \quad (8)$$

For actual computations using Eq. 8, a small but finite time, Δt (not infinitesimal), and the average change in momentum are used as approximations. There is significant uncertainty in evaluating the duration of impact, Δt . SEI/ASCE Standard 7-02 (ASCE 7 2003) recommends a single value of the impact duration, $\Delta t = 0.03$ s. The City and County of Honolulu Building Code (CCH 2000) recommends Δt values for wood construction as 1.0 s, for steel construction as 0.5 s, and for reinforced concrete construction as 0.1 s. The FEMA Coastal Construction Manual (FEMA 55 2000) recommends Δt values between 0.2 and 1.1 s.

Another approach for estimating debris impact forces, which differs from the impulse-momentum approach, was suggested by Haehnel and Daly (2002). The Haehnel and Daly approach is based on a single-degree-of-freedom oscillator. The oscillator consists of a solid body at position $x(t)$ with mass m connected to a support by a spring with constant stiffness k and a dashpot with constant damping coefficient μ . The equation of motion for the single-degree-of-freedom linear oscillator is

$$m\ddot{x} + \mu\dot{x} + kx = 0. \quad (9)$$

The debris collision takes place over a very short duration, so damping is negligible. With no damping term, Eq. 9 reduces to Newton's second law, i.e., $F = ma$. The debris impact force, F , is given by kx , i.e., Hooke's law. By assuming that the impact would occur over one-fourth of a sinusoidal oscillation, the undamped equation of motion, $m\ddot{x} + kx = 0$, yields the maximum debris impact force:

$$F_{\max} = \text{Max}\langle kx \rangle = u\sqrt{km}. \quad (10)$$

Equation 10 involves many simplifying approximations and assumptions without estimating, even qualitatively, the significance of the approximations and assumptions.

The foregoing work clearly demonstrates the immaturity of our present understanding of tsunami-induced debris impact forces. FEMA P646 (2012) recommends that Eq. 10, rather than the impulse-momentum approach, be used to calculate debris impact forces, only because of its simple and "somewhat" rational formulation. FEMA P646 (2012) includes the "added-mass effect" (described in more detail below) by modifying Eq. 10 to

$$F_{\max} = 1.3u_{\max}\sqrt{km(1+c)}, \quad (11)$$

where c is the hydrodynamic mass coefficient that represents the effect of "added mass" and the coefficient 1.3 represents the importance coefficient for critical facilities (i.e., the safety factor). Evidently, manipulating Eq. 11 with multiple empirical correction parameters represents an "engineering struggle" to establish a formula to use in practice.

To understand the added-mass effect, conceptualize a body with mass, m , which is not surrounded by fluid (i.e., in vacuum). To set the body in motion, an external impulse, mu , is required, where u is the velocity. If the same body is immersed in fluid, then the fluid needs an additional impulse, $m_a u$, to create "irrotational flow motion." Accordingly, the total external impulse needed to move the body and the surrounding fluid is $(m + m_a)u$. Hence, the presence of the fluid effectively introduces an "added mass," m_a , to a body of mass m immersed in the fluid. Importantly, the added mass is a correction to the impulse in the form of the impulse equation. Therefore, values of added mass found in the literature are used to correct the total impulse. The parameter c in Eq. 11 does not represent the traditional meaning of the added mass. The functional form of Eq. 11 indicates that the added-mass effect depends on the mass of the body, instead of the fluid mass. In other words, if we have two different objects with the identical shape but different mass, then Eq. 11 indicates that the added-mass effect is more important for the heavier object, which is evidently false. It must be added, however, that, in the case of floating

debris, the mass of the object is the same as the mass of the water displaced by the object; luckily, this fact saves Eq. 11 from being completely incorrect. In summary, there is no rational method available to estimate the debris impact force at the present time, but empirical methods are used to make approximations for design.

Based on a large experimental program including in-air and in-water impact testing, Ko (2012) concluded that attributing the coefficient to the added-mass effect would not accurately describe the debris impact process. Instead, Ko proposed that an experimentally calibrated coefficient could be used to modify the formula for debris impact force given in Eq. 10:

$$F = C_H u \sqrt{km}, \quad (12)$$

where C_H is a hydraulic coefficient, typically ranging between 1.1 and 1.4, depending on the debris material.

The jamming effect of debris on a structure, also known as the damming effect, increases the surface area exposed to the flow, thus increasing the hydrodynamic forces. This force follows after the initial impact force of the debris. FEMA P646 (2012) defines this force using Eq. 3 but replacing the width of the structure with the width of the jammed debris, thus increasing the force.

Remarks

Tsunami hazard prediction for a given locality is a formidable problem with substantial uncertainties. For instance, there are significant uncertainties associated with earthquake prediction methods, fault displacement models, and tsunami generation models, which complicates predicting when and how tsunami waves will attack a shoreline. Even if an offshore design tsunami condition were fully known, it would still be difficult to estimate the tsunami effects on coastal structures, because the tsunami flow conditions are complex and transient, and the onshore terrain, vegetation, and structures are nonuniform and three-dimensional. Notwithstanding these difficulties, engineers must predict the performance of coastal

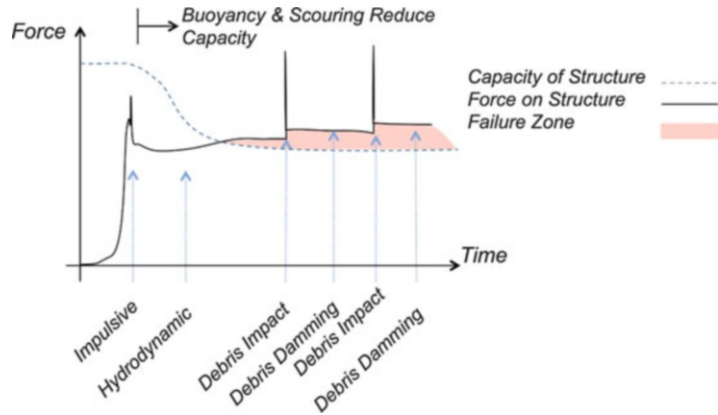
structures during tsunami attack. Most tsunami wave force formulations depend on water wave height and velocity parameters, which are estimated using numerical or analytical solutions with large uncertainties. Accordingly, model input parameters required for force estimation are typically developed using the following approaches: (a) when maximum inundation depth or run-up elevation is the only parameter available or (b) when numerical analysis provides specific height and velocity terms.

For analysis and assessment of structural failures, the total maximum force demands need to be computed. In Fig. 16, an illustrative sketch of a possible force time series is shown as a continuous line. Figure 16 shows the complex combination of tsunami forces and implies the difficulty of estimating a “peak tsunami force.” Figure 16 also shows a dashed line, which represents the capacity of the structure. Note that there is a decrease in structural capacity that can be attributed to the buoyancy force. Recall that the buoyant force reduces the resistance of the structure to sliding and overturning. Buoyancy effects and foundation scouring are very important factors in the evaluation of structural failures. The reduction of the weight of a structure by buoyancy promotes overturning, and the soil acts as a resistance force against overturning. The greater the buoyant force and the greater the scouring, the more susceptible the structures are to failure. Note that tsunami-induced soil scouring and soil instability and the more complete soil–fluid–structure interaction problem were not discussed in depth within this entry. These topics are focuses of current research by the authors (e.g., Yeh and Mason 2014; Yeh et al. 2014) among others and hold great promise for future investigations, as discussed in the next section.

There are several other important factors that should be considered when estimating forces on structures. First, the structural shape and the orientation of the structure with respect to the flow direction affect the value of the drag coefficient. Second, the presence of sacrificial elements, such as breakaway windows and walls in buildings, can provide substantial force alleviation (e.g., Dalrymple and Kriebel 2005; Lukkunaprasit and

Tsunamis Effects in Man-Made Environment,

Fig. 16 An illustrative sketch of tsunami demand and resistance time series for a building structure



Ruangrassamee 2008; Yeh et al. 2013). However, the appropriate “efficiency” of the sacrificial elements is often not known with much certainty. An example illustrating the “efficiency” of breakaway windows and walls is shown in Fig. 17 for two buildings located in Rikuzentakata, Japan. Two flat and thin apartment buildings are oriented with the largest surface area facing the shoreline, yet despite their orientation, the buildings remained standing after the tsunami, most likely due to the force alleviation from the breakaway windows and walls.

In summary, to develop engineering estimates for tsunami effects on coastal structures, we considered the following force classifications: (1) hydrodynamic, (2) hydrostatic, (3) buoyant, (4) surge, and (5) debris impact forces. Among the considered forces, only the hydrodynamic, hydrostatic, and buoyant forces are based on the principles of fluid mechanics. In contrast, estimating surge and debris impact forces currently requires the use of not-well-established empirical methods. The empirical methods are built on engineering judgment, which is based on very limited experience from the past tsunamis. Some of the current tsunami design guidelines, such as FEMA P646 (2012), introduce other categories of tsunami-induced forces, e.g., debris damming forces, uplift forces, and additional gravity loads from retained water on elevated floors. These other forces are similarly difficult to estimate. Because predicting tsunami effects on coastal structures involves large uncertainties, it is crucial to use the predictive formulae correctly. A strong

knowledge of the underlying mechanics is required to balance the involvement of empirical engineering struggles. Lastly, it is emphasized that the correct method to evaluate uplift buoyant forces on buildings with foundation is introduced in this entry. Detailed engineering applications for the effects of buoyant forces can be found in Yeh et al. (2014).

Future Directions

The overarching motivation for studying tsunami effects on coastal structures is to save human lives. Furthermore, our efforts are directed toward the preparation for quick recovery of communities, countries, and regions after the impending disaster. One strategy for protecting coastal areas is to use coastal armoring, such as seawalls and dikes, to either stop or reduce the tsunami impacts. Most coastal armoring is designed for more routine storms, such as hurricanes or typhoons. The unique characteristics of tsunamis, and their relative rareness of occurrence, mean that armoring systems are not usually designed for tsunamis. The east coast of Japan has significant armoring systems, because it is routinely attacked by typhoons and occasionally attacked by tsunamis. Coastal armoring for storm wave protection is rare on the west coast of the United States. The armoring system in Japan was largely overtopped during the 2011 Tsunami but found to be partially effective to mitigate tsunami effects, even if the system was destroyed (e.g., Sato et al. 2014). The



Tsunamis Effects in Man-Made Environment, Fig. 17 Example of breakaway walls in Rikuzentakata, Japan. All other surrounding buildings were toppled and failed

2011 Tsunami was clearly a *beyond-the-design-basis* event and showed us that coastal armoring systems should not be considered as the secured system for all the oceanic natural hazards. Therefore, strategies are needed to minimize loss of life from tsunamis, even if the armoring systems fail. Another strategy is evacuation. Tsunami evacuation strategies have been improved significantly since the 2004 Indian Ocean Tsunami. Evacuation is effective for coastal areas that have higher elevations nearby. Nonetheless, in many coastal areas, the lack of nearby higher ground and the potential traffic volumes make evacuation a less obvious choice for reducing tsunami casualties. In such cases, building tsunami-resistant structures should be an option so that people have an option to escape themselves to the higher elevation of the structures as the tsunami passes. There is a great future potential to construct (or retrofit) tsunami-resistant buildings at key locations for the use of vertical evacuations. Evidently, engineering design bases must be established for tsunami-resistant buildings.

The 2011 Great East Japan Tsunami was devastating and there are many lessons learned. The 2011 Tsunami attacked a well-developed and well-prepared nation. Prior to the event, Japanese scientists and engineers had key instrumentation deployed, and the well-educated general public was able to capture high-quality video footage as well as photographs. The dataset from the 2011 Tsunami is the best, both in quality and quantity, of all the previous tsunami datasets. As the datasets of the 2011 Tsunami are investigated, more guidance for tsunami effects on structures will be developed; therefore, we hope to prepare for the future megatsunami hazards more effectively and reduce damage from a similar disaster.

Finally, when we study tsunami effects on coastal structures, it is essential to consider the interactions of fluid, soil, and structure as a system. For instance, certain tsunami conditions (e.g., certain flow depth and flow velocity combinations) can lead to significant soil instability, which results in significant formations of scour (e.g., Sumer and Fredsøe 2002) or leads to

momentary liquefaction (e.g., Yeh and Mason 2014). Such soil instability can induce foundation instability and subsequent structure failure. Although this entry largely focused on tsunami-induced forces on structures, moving forward, it is important to examine soil instability to solve the complete fluid–soil–structure interaction problem. Together with substantial uncertainty in the prediction of tsunami hazard itself, such multiphase and nonlinear problems are evidently formidable.

Bibliography

- Abe K, Tsuji Y, Imamura F, Katao H, Yohihisa L, Satake K, Bourgeois J, Noguera E, Estrada F (1993) Field survey of the Nicaragua earthquake and tsunami of September 2, 1992 (in Japanese). *Bull Earthq Res Inst Univ Tokyo* 68: 23–70
- Arnason H (2005) Interactions between an incident bore and a free-standing coastal structure. PhD thesis, University of Washington, Seattle, 172 pp
- Arnason H, Petroff C, Yeh H (2009) Tsunami bore impingement onto a vertical column. *J Disaster Res* 4(6):391–403
- ASCE 7 (2003) Minimum design loads for buildings and other structures. *SEI/ASCE 7-02*, 376 pp
- Batchelor GK (1967) An introduction to fluid mechanics. Cambridge University Press, Cambridge, 615 pp
- Biot MA (1941) General theory of three-dimensional consolidation. *J Appl Phys* 12:155–164
- CCH (2000) Department of planning and permitting of Honolulu Hawai'i. City and County of Honolulu Building Code. Chapter 16, Article 11
- Cross RH (1967) Tsunami surge forces, journal of waterways and harbor division. *ASCE* 93(4):201–231
- Cumberbatch E (1960) The impact of a water wedge on a wall. *J Fluid Mech* 7:353–374
- Dalrymple RA, Kriebel DL (2005) Lessons in engineering from the tsunami in Thailand. *Bridge* 35(2):4–13
- Dames, Moore (1980) Design and construction standards for residential construction in tsunami-prone areas in Hawaii, prepared for the Federal Emergency Management Agency
- Disaster Countermeasures Office (2011) About the accident of Fukushima nuclear power plants no. 1 and 2 in 2011, 166 p (in Japanese)
- FEMA 55 (2000) Coastal construction manual. Federal Emergency Management Agency, Washington, DC
- FEMA P646 (2012) Guidelines for design of structures for vertical evacuation from tsunamis. Federal Emergency Management Agency, Washington, DC
- Haehnel RB, Daly SF (2002) Maximum impact force of woody debris on floodplain structures. Technical report: ERDC/CRREL TR-02-2, US Army Corps of Engineers, 40 pp
- Holtz RD, Kovacs WD (1981) An introduction to geotechnical engineering. Prentice Hall, Englewood Cliffs, 733 pp
- Ko H (2012) Hydraulic experiments on impact forces from tsunami-driven debris. MS thesis, Oregon State University, 193 pp
- Lamb H (1960) *Statics*. Cambridge University Press, Cambridge, England, 361 pp
- Lay T, Wallace T (1995) *Modern global seismology*. Academic, San Diego, 521 pp
- Lukkunaprasit P, Ruangrassamee A (2008) Building damage in Thailand in the 2004 Indian Ocean tsunami and clues for tsunami-resistant design. *IES J Part A Civ Struct Eng* 1(1):17–30
- Lynett P, Borrero J, Weiss R, Son S, Greer D, Renteria W (2012) Observations and modeling of tsunami-induced currents in ports and harbors. *Earth Planet Sci Lett* 327(328):68–74
- Mansinha L, Smylie D (1971) The displacement fields of inclined faults. *Bull Seismol Soc Am* 61:1433–1440
- Miller C, Cabbage A, Dorman D, Grobe J, Holahan G, Sanfilippo N (2011) Recommendations for enhancing reactor safety in the 21st Century. The near-term task force review of insights from the Fukushima Dai-Ichi accident. U.S. Nuclear Regulatory Commission, Rockville, 83 pp
- Mori N, Takahashi T, The Tohoku Earthquake Tsunami Joint Survey Group (2012) Nationwide post event survey and analysis of the 2011 Tohoku earthquake tsunami. *Coast Eng J* 54:4, 27 p
- Nakao H, Sato S, Yeh H (2012) Laboratory study on destruction mechanisms of coastal dyke due to overflowing tsunami. *Proc Coast Eng* 68:281–285
- National Tsunami Hazard Mitigation Program (2014) <http://nws.weather.gov/nthmp/>. Accessed 20 Aug 2014
- Okada Y (1985) Surface deformation due to shear and tensile faults in a half-space. *Bull Bull Seismol Soc Am* 75:1135–1154
- Okal EA, Fritz HM, Raad PE, Synolakis CE, Al-Shijbi Y, Al-Saifi M (2006) Oman field survey after the December 2004 Indian Ocean tsunami. *Earthq Spectra* 22:S203–S218
- Paczkowski K, Riggs R, Robertson I (2011) Bore impact upon vertical wall and water-driven, high-mass, low-velocity debris impact, research report UHM/CEE/11-05. University of Hawaii
- Ramsden JD (1993) Tsunamis: forces on a vertical wall caused by long waves, bores, and surges on a dry bed. Report no. KH-R-54. California Institute of Technology, Pasadena
- Sato S, Okayasu A, Yeh H, Fritz HM, Tajima Y, Shimozono T (2014) Delayed survey of the 2011 Tohoku Tsunami in the former exclusion zone in Minami-Soma, Fukushima Prefecture. *Pure Appl Geophys*. doi:10.1007/s00024-014-0809-8
- Shuto N (1994) Building damages due to the Hokkaido Nansei-Okai Earthquake and Tsunami. Tsunami Engineering Technical report no. 11, Tohoku University, pp 11–28. (in Japanese)

- Sumer BM, Fredsøe J (2002) The mechanics of scour in the marine environment. World Scientific, Singapore
- wTakahashi R, Hatori T (1961) A summary report on the Chilean Tsunami of May 1960. Report on the Chilean Tsunami, Field Investigation Committee for Chilean Tsunami
- Takahashi T, Imamura F, Shuto N (1992) Research on flows and bathymetry variations by tsunami: the case of Kesennuma Bay, Japan, due to the 1960 Chilean Tsunami (in Japanese). Tsunami engineering technical report no. 9, Tohoku University, pp 185–201
- Terzaghi K (1943) Theoretical soil mechanics, John Wiley and Sons, New York, pp. 510
- United States Geologic Survey (2014) <http://earthquake.usgs.gov/earthquakes/eqarchives/year/eqstats.php>. Accessed 20 Aug 2014
- Wachtendorf T, Kendra JM, Rodriguez H, Trainor J (2006) The social impacts and consequences of the December 2004 Indian ocean tsunami: observations from India and Sri Lanka. *Earthq Spectra* 22(S3): 693–714
- Yeh H (2006) Maximum fluid forces in the tsunami runup zone. *J Waterw Port Coast Ocean Eng* 132:496–500
- Yeh H (2007) Design tsunami forces for onshore structures. *J Disaster Res* 2:531–536
- Yeh H, Mason HB (2014) Sediment response to tsunami loading: mechanisms and estimates. *Geotechnique* 64(2):131–143
- Yeh H, Barbosa A, Ko H, Cawley J (2014) Tsunami loadings on structures: review and analysis. In: Proceedings of the 34th conference on coastal engineering, Seoul, Korea
- Yeh H, Mok K-M (1990) On turbulence in bores. *Phys Fluids* 2:821–828
- Yeh H, Liu P, Synolakis C (1996) Long-wave runup models. World Scientific, Singapore, 403 pp
- Yeh H, Francis M, Peterson C, Katada T, Latha G, Chadha RK, Singh JP, Raghuraman G (2007) Effects of the 2004 Great Sumatra Tsunami: Southeast Indian Coast. *J Waterw Port Coast Ocean Eng* 133:382–400
- Yeh H, Sato S, Tajima Y (2013) The 11 March 2011 East Japan earthquake and tsunami: tsunami effects on coastal infrastructure and buildings. *Pure Appl Geophys* 170(6-8):1019–1031
- Zen K, Yamazaki H (1990) Mechanism of wave-induced liquefaction and densification in seabed. *Soils Found* 30(4):90–104
- Zen K, Yamazaki H (1991) Field observation and analysis of wave-induced liquefaction in seabed. *Soil Found* 31(4):161–179
- Journal Papers**
- Ewing L (2011) The Tohoku Tsunami of March 11, 2011: a preliminary report on effects to the California coast and planning implications. A report to Coastal Commissioners, 40 pp. http://www.coastal.ca.gov/energy/tsunami/CCC_Tohoku_Tsunami_Report.pdf
- Geist EL (1999) Local tsunamis and earthquake source parameters. *Adv Geophys* 39:117–209
- Gokon H, Koshimura S (2012) Mapping of building damage of the 2011 Tohoku earthquake tsunami in Miyagi Prefecture, Coast Eng J 54:1250006 12 p
- Hughes SA (2003) Wave momentum flux parameter for coastal structure design. Report no. ERDC/CHL CHETN-III-67, US Army Corps of Engineers, Vicksburg, 12 pp
- Ikeno M, Tanaka Y (2003) Experimental study on impulse force of drift body and tsunami running up to land. *Proc Coast Eng Jpn Soc Civ Eng* 50:721–725 (in Japanese)
- Ikeno M, Mori N, Tanaka Y (2001) Experimental study on tsunami force and impulsive force by a drifter under breaking bore like tsunamis. *Proc Coast Eng Jpn Soc Civ Eng* 48:846–850 (in Japanese)
- Kato F, Suwa Y, Watanabe K, Hatogai S (2012) Mechanisms of coastal dike failure induced by the Great East Japan earthquake and tsunami. In: Proceedings of the 33rd Conference on Coastal Engineering, Santander, Spain
- Kawashima K (2012) Damage of bridges due to the 2011 Great East Japan Earthquake. *Journal of Japan Association for Earthquake Engineering* 12(4):319–338
- Koshimura S, Namegaya Y, Yanagisawa H (2009) Tsunami fragility – a new measure to identify tsunami damage. *J Disaster Res* 4(6):479–488
- Lukkunaprasit P, Thanasisathit N, Yeh H (2009) Experimental verification of FEMA P646 tsunami loading. *J Disaster Res* 4(6):410–418
- Matsutomi H (1999) A practical formula for estimating impulsive force due to driftwoods and variation features of the impulsive force. *Proc Jpn Soc Civ Eng* 621: 111–127 (in Japanese)
- Matsutomi H, Shuto N (1994) Tsunami engineering technical report no.11, DCRC, Tohoku University, pp 29–32
- Murakami H, Fraser S, Leonard GS, Matsuo I (2012) A field study on conditions and roles of tsunami evacuation buildings in the 2011 Tohoku Pacific earthquake and tsunami. In: Proceedings of the 9th international conference of urban earthquake engineering, Tokyo, pp 89–95
- Pacheco KH, Robertson IN (2005) Evaluation of tsunami loads and their effect on concrete buildings, University of Hawaii research report UHM/CEE/05-06, pp 189, 207
- PG&E (2010) Methodology for probabilistic tsunami hazard analysis: trial application for the Diablo Canyon power plant site. PEER workshop on tsunami hazard analyses for engineering design parameters, Berkeley
- Raskin J, Wang Y, Boyer M, Fiez T, Moncada J, Yu K, Yeh H (2011) An evacuation building project for Cascadia earthquake and tsunamis. *Obras Y Proyectos Rev Ingenieria Civ* (translation: Work and Projects: Civil Engineering Magazine) 9:11–22
- Reese S, Cousins WJ, Power WL, Palmer NG, Tejakusuma IG, Nurgrahadi S (2007) Tsunami vulnerability of buildings and people in South Java – field observation after the July 2006 Java tsunami. *Nat Hazards Earth Syst Sci* 7:573–589

- Robertson IN, Riggs RH, Yim S, Young YL (2006) Lessons from Katrina, *Civil Engineering*. Am Soc Civ Eng 76(4):56–63
- Ruangrassamee A, Yanagisawa H, Foytong P, Lukkunaprasit P, Koshimura S, Imamura F (2006) Investigation of tsunami-induced damage and fragility of buildings in Thailand after the December 2004 Indian Ocean Tsunami. *Earthq Spectra* 22(S3): 377–401
- Saatcioglu M, Ghobarah A, Nistor I (2005). Reconnaissance Rep. on the December 26, 2004 Sumatra Earthquake and Tsunami, Canadian Association for Earthquake Engineering, Ottawa, Canada
- Saatcioglu M, Ghobarah A, Nistor I (2006) Performance of structures in Indonesia during the December 2004 Great Sumatra earthquake and Indian Ocean tsunami. *Earthq Spectra* 22:S3
- Tinti S, Tonini R, Bressan L, Armigliato A, Gargi A, Guillande R, Valencia N, Scheer S (2011) Handbook of tsunami hazard and damage scenarios, SCHEMA Project, JRC Scientific and technical reports, EUR 24691 EN
- Tonkin S, Yeh H, Kato F, Sato S (2003) Tsunami scour around a cylinder. *J Fluid Mech* 496:165–192
- Yeh H (2008) Closure to maximum fluid forces in the tsunami runup zone. *J Waterw Ports Coast Ocean Eng* 134:200–201
- Yeh H, Robertson I, Preuss J (2005) Development of design guidelines for structures that serve as tsunami vertical evacuation sites, open file report 2005-4, Washington Division of Geology and Earth Resources, State of Washington (contract 52-AB-NR- 200051), Olympia
- ## Books and Reports
- Bernard RN, Robinson AR (eds) (2009) *Tsunamis: the sea*, vol 15. Harvard University Press, Cambridge, MA, 450 pp
- Camfield F (1980) *Tsunami engineering*, Coastal Engineering Research Center, US Army Corps of Engineers, Special Report (SR-6), 222 pp
- Fukuyama H, Okoshi T (2005) Introduction of the tsunami resisting design method for buildings proposed by the building center of Japan. Building Center of Japan, Tokyo
- Henderson FM (1966) *Open channel flow*. Macmillan, New York, 522 pp
- Iwan WD (ed) (2006) *Earthquake spectra, special issue III*, vol 22: the Great Sumatra earthquakes and Indian Ocean tsunamis of 26 December 2004 and 28 March 2005 reconnaissance report, EERI, 900 pp
- Murty TS (1977) *Seismic sea waves tsunamis*, vol 198, Bulletin. Fisheries and Marine Service, Ottawa, 337 pp
- National Research Council (2011) *Tsunami warning and preparedness*. The National Academies Press, Washington, DC, 284 pp
- Wiegel RL (2005) *Tsunami information sources*. Hydraulic Engineering Laboratory technical report UCB/HEL 2005-1, University of California, Berkeley
- Wiegel RL (2006a) *Tsunami information sources*, part 2. Hydraulic Engineering Laboratory technical report UCB/HEL 2006-1, University of California, Berkeley
- Wiegel RL (2006b) *Tsunami information sources*, part 3. Hydraulic Engineering Laboratory technical report UCB/HEL 2006-3, University of California, Berkeley
- Wiegel RL (2008) *Tsunami information sources*, part 4. Hydraulic Engineering Laboratory technical report UCB/HEL 2008-1, University of California, Berkeley

Tsunami Hazard and Risk Assessment on the Global Scale

F. Løvholt¹, J. Griffin² and
M. A. Salgado-Gálvez³

¹Norwegian Geotechnical Institute (NGI), Oslo, Norway

²Geoscience Australia, Canberra, ACT, Australia

³Centre Internacional de Metodes Numerics en Enginyeria (CIMNE), Barcelona, Spain

Article Outline

Glossary

Definition of Subject

Introduction

Review of Previous Tsunami Hazard Assessments

Global-Scale Tsunami Hazard and Risk Models

Global Tsunami Hazard and Loss Estimation

A Note About Simplifying Assumptions and Uncertainties

Future Directions

Bibliography

Glossary

Digital Elevation model (DEM) A digital model of the surface of topography or elevation usually defined as height above or below mean sea level.

Exposure (Elements at risk) Assets and/or people that are located in areas that may be impacted by a hazard.

Hazard A hazard is a man-made or natural phenomenon with the potential to be harmful to population society and the environment. Hazard assessment refers to efforts to characterise the location intensity and/or temporal probability of hazard events.

Hazard curve Distribution of a hazard intensity as a function of the temporal probability of exceedance.

Loss exceedance curve Distribution of a measure of loss (e.g., economic, fatality) and the probability of exceedance within a given period of time.

Loss return period The inverse value of the loss exceedance rate. It represents the mean time interval on which the selected loss level is expected to occur at least once.

Magnitude Frequency Distribution (MFD) A relation between a measure of the size or intensity of a phenomena (for example an earthquake magnitude) and the probability of being observed within a given time interval.

Mortality rate The probability of a fatality given a tsunami metric.

Physical vulnerability Potential degree of structural and non-structural damage of an exposed element when subjected to a hazard event. For fully probabilistic risk assessments the vulnerability is represented by vulnerability functions allow calculating the monetary losses associated to them.

Probabilistic Tsunami Hazard Assessment (PTHA) A probabilistic description of the size and occurrence frequency quantified in terms of an exceedance probability for the values of a physical intensity or tsunami metric (e.g., flow depth).

Probabilistic Tsunami Risk Assessment (PTRA) A probabilistic description of the size and exceedance frequency of tsunamigenic losses.

Probability density function (PDF) The probability density function is the derivative of the total (cumulative) probability and gives the relative likelihood of all possible random variables. The integral over all possible random variables in the PDF equals unity.

Probable Maximum Loss (PML) A probabilistic risk metric that takes account of the expected size of maximum losses associated

to extreme events. It is always associated to a mean return period.

Tsunami risk A measure of the magnitude and frequency of losses due to tsunamis. It is calculated as the convolution between the hazard and the physical vulnerability of the exposed assets. Quantitatively the Risk is a function of (Hazard, Exposure and Vulnerability). Tsunami risk can be expressed in terms of loss exceedance rates, average annual losses and probable maximum losses.

Vulnerability (1) The degree of loss to a given element at risk or set of such elements resulting from a hazard event of a given magnitude or intensity, usually expressed on a scale from 0 (no loss) to 1 (total loss). (2) Degree of damage caused by various levels of loading. Vulnerability may be calculated in a probabilistic or deterministic way for a single structure or a set of structures.

Definition of Subject

Tsunamis are infrequent events with the power to cause massive loss of life, large economic losses, and cascading effects such as destruction of critical facilities. The recurrence of truly disastrous tsunamis at any location may range from hundreds to even thousands of years. To manage the risk from these events, scientists use hazard and risk assessment to better understand the threat. Hazard assessment typically involves quantifying the temporal probability of a tsunami metric (e.g., run-up height at a coastal location) being exceeded within a given time frame. Risk assessment quantifies the probability of damage and loss to exposed assets and population by integrating the results of the hazard assessment with the vulnerability of the exposed elements to the given tsunami metric. Tsunami hazard and risk analysis cannot exploit observational data as extensively as more frequent hazards. Instead, numerical models are used to quantify the magnitude and frequency of tsunami run-up and inundation at coastal locations within the framework of Probabilistic Tsunami Hazard Assessment (PTHA). Probabilistic tsunami hazard can be convolved

with vulnerability models to quantify the probability of damages and losses to exposed elements, an analysis termed probabilistic tsunami risk assessment (PTRA). We present herein the results of the most up-to-date probabilistic analysis of global tsunami hazard and risk, at present based solely on large earthquake sources. Being a global analysis comprising the most relevant tsunami sources and probabilities, the result of this analysis provides an overview of the global tsunami risk state of affairs. In practice, global analyses such as that presented here should be complemented by more detailed, local studies that inform local-level disaster management decision-making.

Introduction

The 2004 Indian Ocean tsunami and the 2011 Tohoku tsunami have changed our understanding of how to deal with high-impact–low-probability events (Stein and Okal 2007; Kagan and Jackson 2013). In the aftermath of these events, quantification of tsunami hazard around the world has received substantially increased attention. The Hyogo Framework for Action (HFA) was established by the United Nations International Strategy for Disaster Risk Reduction (UN-ISDR) in order to reduce disaster loss, with the 2004 Indian Ocean tsunami providing a catalyst for strong global action. Recently the HFA was followed by the Sendai Framework for Disaster Risk Reduction (SFDRR) which also was marked by the occurrence of the 2011 Tohoku earthquake and tsunami. A sound understanding of disaster risk due to different natural hazards at a global scale, and how this changes over time, underpins implementation of the goals of these agreements. Therefore an integrated worldwide study was implemented and published in several biannual Global Assessment Reports on Disaster Risk Reduction (GAR) published by UN-ISDR from 2009 to 2015 (UN-ISDR 2015).

A number of tsunami hazard assessments have been carried out in different parts of the world, focusing often on the local and regional effects of tsunamis. These studies are usually rather

detailed and often aim to develop high-resolution local inundation maps. However, as outlined above, there is also a need to quantify the potential effects of tsunamis at a wider geographic scale. While global tsunami risk analysis follows a similar methodology to that applied in local studies, some adaption of the methodology is necessary to make the analysis feasible.

The first global-scale tsunami hazard assessment employed a scenario-based, rather than probabilistic, method (Løvholt et al. 2012a). In addition to quantifying the hazard, it also provided an analysis of the population and economic assets exposed to the modeled tsunami scenarios. This study was rather simple since it included a limited compilation of sources supplemented with literature findings, where available, and was also not quite global as certain areas were not included. A complete global hazard and population exposure analysis was included in a follow-up study (Løvholt et al. 2014a) where some preliminary steps were taken toward a fully probabilistic assessment. Furthermore, this analysis demonstrated the exposure of critical facilities, such as nuclear power plants, around the world. While these previous studies targeted a single return period (500 years), at present we now have a more fully developed analysis using a fully probabilistic methodology for assessment of both tsunami hazard and risk.

In the present entry, we first provide a brief review of previous efforts to assess tsunami hazard, separated into the different methodologies (empirical run-up data, scenario-based methods, and probabilistic methods), while also discussing some of the limitations of them. Secondly, we provide a review of the hazard, vulnerability, exposure, and risk assessment methodology, with emphasis on the adaptation for a global-scale study. Next, we present examples of a global tsunami hazard and risk analysis, presenting the results from both previous deterministic studies (Løvholt et al. 2014a), as well as fully probabilistic analysis (Horspool et al. 2014 and work within this entry). To this end, various sources of uncertainties are important, and they are given considerable consideration. While the analysis is constrained to tsunamis generated by

earthquakes, we also include some discussion related to nonseismic sources such as landslides and volcano flank collapses.

Review of Previous Tsunami Hazard Assessments

Use of Empirical Run-Up Data

Available historical tsunami inundation and run-up data may be utilized to construct tsunami hazard curves. Such run-up data may be taken from available online catalogues such as the NGDC (2015) and Tsunami Laboratory Novobirsk (2015) databases. An example of such a study is the statistical analysis of tsunami run-up data on a stretch of coastline covering northern Chile and Peru, conducted by Kulikov et al. (2005). Another example is the one by Burroughs and Tebbens (2005) for Honshu, Japan. In those two analyses, maximum run-up data were fitted to a lognormal cumulative statistical distribution and a power law, respectively. In this way, a statistical model for the temporal occurrence of a tsunami of a certain run-up height at any point of the coastline was established. While useful for a first-order approach, there are limitations by only using historical records for assessing the tsunami hazard, and therefore for the use of those results in subsequent tsunami risk analysis. First, it is limited by the available historical record, which may be short, in this case 100–200 years, relative to the recurrence of large tsunamis. Obviously and as also demonstrated indirectly by Geist and Parsons (2006), the statistical behavior of the run-up probability may be subjected to biases when used beyond the observational record as the low probability tail of the distribution may not be fitted. We note, however, that availability of paleotsunami data can supplement the historical record extending the length of the observational record, as demonstrated in a similar study by Kaistrenko et al. (2003) for the Kamchatka east coast. Secondly, a statistical model of this kind may also only provide maximum values, as these are what tend to be recorded, and not the distribution of the hazard intensities along the coastline. The latter is not

only important for hazard assessment but also for providing useful input to a comprehensive risk assessment.

Use of Scenario-Based Methods

The simplest and most common way of performing tsunami hazard analysis is through simulation of tsunami scenarios. Often, they are described as “credible worst case scenarios,” defined as the maximum intensities that could be expected associated to the occurrence of them. The scenarios may or may not have associated a temporal probability, and in most cases, the assigned probabilities are poorly constrained. Design of such scenarios may use, for instance, the size and location of previous tsunami events (e.g., Barkan et al. 2009), regional seismicity characteristics (Løvholt et al. 2012b), constraints given by tectonic structures (e.g., as demonstrated by Basili et al. (2013)), use of earthquake scaling laws (as in Løvholt et al. (2006)), or a combination of the above. While there is no established standard way for designing or selecting the scenarios, this approach has been popular due to its simplicity, robustness, and ease for communication. Particularly in the immediate aftermath of the 2004 Indian Ocean tsunami, the scenario-based method has been used extensively for first-order hazard assessments, for instance, in Australia (Burbidge and Cummins 2007), South and South East Asia (Løvholt et al. 2006, 2012a; Borrero et al. 2006; Okal and Synolakis 2008; Okal et al. 2011), the western United States and South America (Legg et al. 2004; Okal et al. 2006; Venturato et al. 2007; Brizuela et al. 2014), the Caribbean (Mercado 2001; Harbitz et al. 2012), the North Atlantic (Barkan et al. 2009; Grilli et al. 2010), and the Mediterranean (Tinti and Armigliato 2003; Hebert et al. 2005; Roger and Hebert 2008; Lorito et al. 2008). Despite its popularity, it is noted that the method is highly sensitive to the scientific judgment involved when selecting the model parameters, which makes model comparisons less straightforward. This subjectivity may be of first-order importance in areas of complex tectonics such as the Mediterranean (Tinti and Armigliato 2003) and Eastern Indonesia

(Løvholt et al. 2012b) and to an even larger extent for landslide sources (Harbitz et al. 2014). Subjective choices are also involved in more sophisticated probabilistic methods such as PTHA (see below for a description). However, the PTHA method provides a framework to understand how these choices affect the end result and to include and weight alternative models where there is uncertainty and disagreement.

Use of Probabilistic Tsunami Hazard Assessment

The use of PTHA has been increasingly popular following the landmark development of Geist and Parsons (2006). PTHA was developed by adapting the well-established probabilistic seismic hazard assessment (PSHA) method (Esteve 1967; Cornell 1968) and has gradually established itself as a common methodology for tsunami hazard analysis. Examples of applications include Australia (Burbidge et al. 2008a, b), South and South East Asia (Heidarzadeh and Kijko 2011; Suppasri et al. 2012; Horspool et al. 2014), the Southern Pacific (Lane et al. 2013), East Asia (Annaka et al. 2007; Liu et al. 2007), the western United States (Gonzalez et al. 2009), the Caribbean (Parsons and Geist 2009), the North Atlantic (Omira et al. 2014), and the Mediterranean (Sørensen et al. 2012; Lorito et al. 2015). In the next sections, we explore a version of the PTHA employed towards a global tsunami hazard analysis further.

Global-Scale Tsunami Hazard and Risk Models

Overview of the Global Probabilistic Tsunami Hazard and Risk Method

The Probabilistic Tsunami Hazard Assessment (PTHA) approach describes the probability of exceedance for a given metric of tsunami intensity. The first development of PTHA was the theoretical model of Lin and Tung (1982). Modern applications of PTHA are, however, based on the developments of Geist and Parsons (2006), who quantified the probability of off-shore tsunami amplitudes as the tsunami

metric. Modifications of their approach has been followed by others (e.g., Thio et al. 2010; Horspool et al. 2014), as calculation of offshore tsunami heights can be done with linear tsunami propagation models. This makes feasible the calculation of a high number (at present more than 10^5) of possible future events and their associated probabilities of occurrence. However, a significant limitation of this approach is that neither the extent nor the depth of tsunami inundation is calculated and, therefore, risk also cannot be calculated. Some recent site-specific studies have partially addressed this limitation by using limited sets of events derived from the PTHA to simulate tsunami inundation using detailed nonlinear shallow water (NLSW) models (Gonzalez et al. 2009; Lorito et al. 2015). NLSW models are more computationally intensive than linear propagation models, and, more importantly, they require high resolution and accuracy elevation data to robustly simulate inundation (Griffin et al. 2015).

The present discussion focuses on calculating tsunami hazard and risk assessment over large geographic regions. Hence, some hybrid method that conveys the offshore near-shore amplitudes to an onshore tsunami metric such as the run-up height is needed. In addition, the method needs to be able to provide rough inundation maps for integration with the exposure database. For a global study such as this one, taking into account that a large number of events are needed, computationally faster but more approximate methods than solving the NLSW over high-resolution elevation models must be employed. While the PTHA implemented for this analysis follows closely the conventional method of Horspool et al. (2014), our present methodology includes some additional adaptation to estimate an onshore hazard metric, as well as the subsequent risk analysis. The PTHA and associated risk analysis, labeled PTRa, can then be briefly summarized as follows:

1. Define the tsunami sources (earthquake faults) to be included in the analysis.
2. For each fault zone, discretize the fault into smaller subfaults (unit sources).

3. For each source develop a synthetic earthquake catalogue based on a magnitude frequency distribution (MFD) of choice (e.g., Gutenberg-Richter or Characteristic). The recurrence model relates magnitude to the frequency of occurrence and is controlled by the slip rate (convergence rate) and coupling of the source. Each synthetic event is a linear combination of adjacent subfaults with the number of subfaults and amount of slip dependent on the moment magnitude of the event. The same slip is applied to every subfault. For each magnitude, the full set of synthetic sources will cover the whole geometry of the subduction zones.
4. For each subfault, calculate the seafloor deformation for unit (1 m) slip and use this as the initial condition disturbance of the sea surface.
5. Propagate the resulting tsunami from source to hazard points offshore from the coastline of interest at a reference depth, using linear wave theory.
6. For each event in the synthetic earthquake catalogue, estimate the maximum water level at the near-shore hazard point by summing the waves from all the individual subfaults that make up that event, and then scaling by the amount of slip for that event.
7. Amplify the near-shore maximum tsunami water level to estimate maximum run-up height using amplification factors calculated from nonlinear modeling of tsunami over idealized bathymetry profiles.
8. Estimate the inundation area for the event by finding all coastal areas below the estimated run-up height using a global digital elevation model (DEM).
9. Overlay the inundation zones with a dataset containing the exposed elements at risk.
10. Assign a vulnerability function to each exposed asset. In this case, the vulnerability functions relate the mean damage ratio to the local flow depth of a tsunami event.
11. Compute the Loss Exceedance Curve (LEC), by convoluting the hazard and the physical vulnerability considering all feasible events and exposure databases. From this risk

metric, other probabilistic loss metrics such as the Annual Average Loss (AAL) or the Probable Maximum Loss (PML) can be derived.

Earthquake Source Modeling

As the largest tsunamigenic earthquakes occur on subduction zones, these tectonic features are primarily used in deriving the earthquake source model. In a few regions, including the Mediterranean, eastern Indonesia, and North-East Atlantic regions, nonsubduction sources are considered regionally as significant drivers of hazard and must also be included in tsunami hazard analysis. Global tectonic models such as those of Bird (2003), the Slab 1.0 model (Hayes et al. 2012) and the Global Earthquake Model (GEM) Faulted Earth Subduction Characterisation Project (Berryman et al. 2013) are used to derive the geometry of major faults (Fig. 1) and their tectonic slip-rate (i.e., the rate of movement across the fault). The slip-rate, along with estimates of the maximum possible magnitude for a particular fault and the ratio of large-to-small earthquakes, can be used to construct a magnitude-frequency distribution for earthquakes on a particular fault. Regional and national studies (e.g., Burbidge et al. 2008a, b; Parsons and Geist 2009; Matias et al. 2013; Horspool et al. 2014; Omira et al. 2014) may involve a more detailed fault model and include nonsubduction sources that may be locally significant.

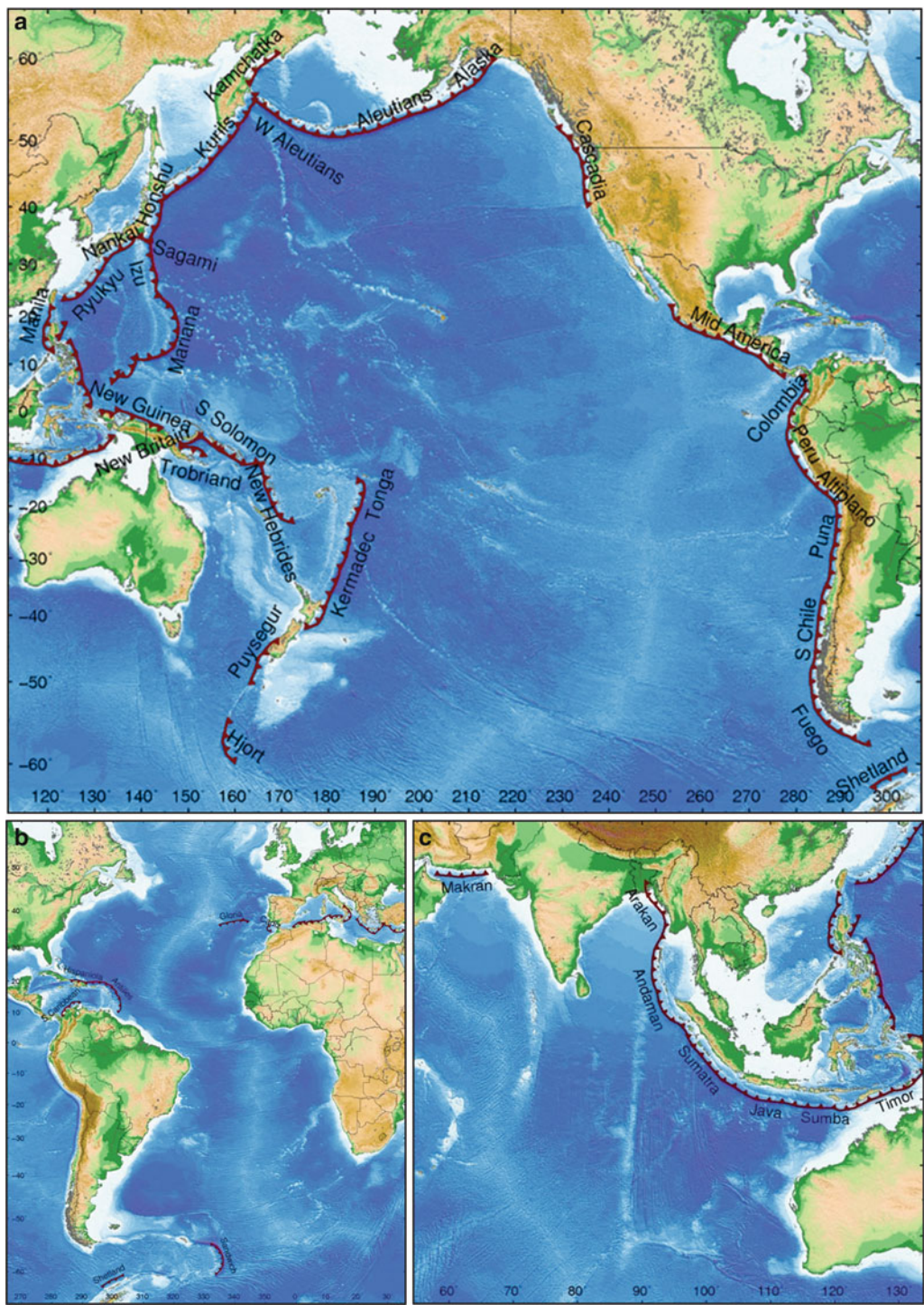
The maximum depth of the seismogenic zone is taken from the GEM database (Berryman et al. 2013). Ruptures may propagate to the trench in all source regions except the Hellenic Arc, as in this region the trench is located significantly further south (near the north coast of Africa) than the expected up-dip limit of the active megathrust.

Once the geometry of source fault planes has been determined, the fault interface is discretized into “small” (in this case, 5×5 km) subfaults resulting in a high-resolution definition of the fault structure. The 5×5 km subfaults are then grouped into larger 100×50 km “large” subfaults. For each large subfault the surface

deformation resulting from a unit 1 m of slip on each of its component small subfaults is calculated and aggregated; this deformation is then used as an initial condition to the tsunami propagation model. This process is illustrated in Fig. 2. Through linear combinations, the large subfaults then form the basis of the probabilistic analysis to create the synthetic event database. This approach allows resolution of nonlinear subduction interface geometries while maintaining a coarser grid size for the tsunami propagation calculations and reducing the complexity of the probabilistic calculations.

The earthquake event database is constructed by iterating through each magnitude in the magnitude-frequency distribution (in moment magnitude steps of 0.1) and calculating the rupture dimensions and amount of slip based on a scaling law (Strasser et al. 2010). The rupture is then iteratively moved across the fault until that magnitude has occurred on every possible location (i.e., every possible contiguous combination of subfaults) within the fault dimensions. This iterative process ensures that all magnitudes could occur at any possible location on the fault plane. For each event its probability is then the probability of that magnitude occurring on the fault divided by the number of earthquake events of that magnitude in the synthetic event database.

The deformation of the earth’s surface resulting from slip on a fault can be calculated using models of the properties of the overlying crust. For homogeneous crustal properties, an exact analytical solution exists (Okada 1985). The real structure of the crust can be better approximated by a series of layers with different crustal properties (Wang et al. 2003). This approach is taken here, using a generic layered earth model based on the Crust 1.0 continental arc structure (Table 1, Laske et al. 2012). It is noted that the Crust 1.0 model includes a variety of relevant crustal structure classes, including forearcs, islands arcs, continental arcs, continental margins, and oceanic crust that overlap with the location of global subduction zones. Uniform application of the continental arc structure everywhere is a reasonable approximation because



Tsunami Hazard and Risk Assessment on the Global Scale, Fig. 1 Traces (in red) of major tsunamigenic faults for (a) the Pacific Ocean region; (b) the Atlantic Ocean region; and (c) the Indian Ocean region (additional

sources included in the analysis for eastern Indonesia and the Philippines are not shown at this scale, see NGI and Geoscience Australia (2015), for details)

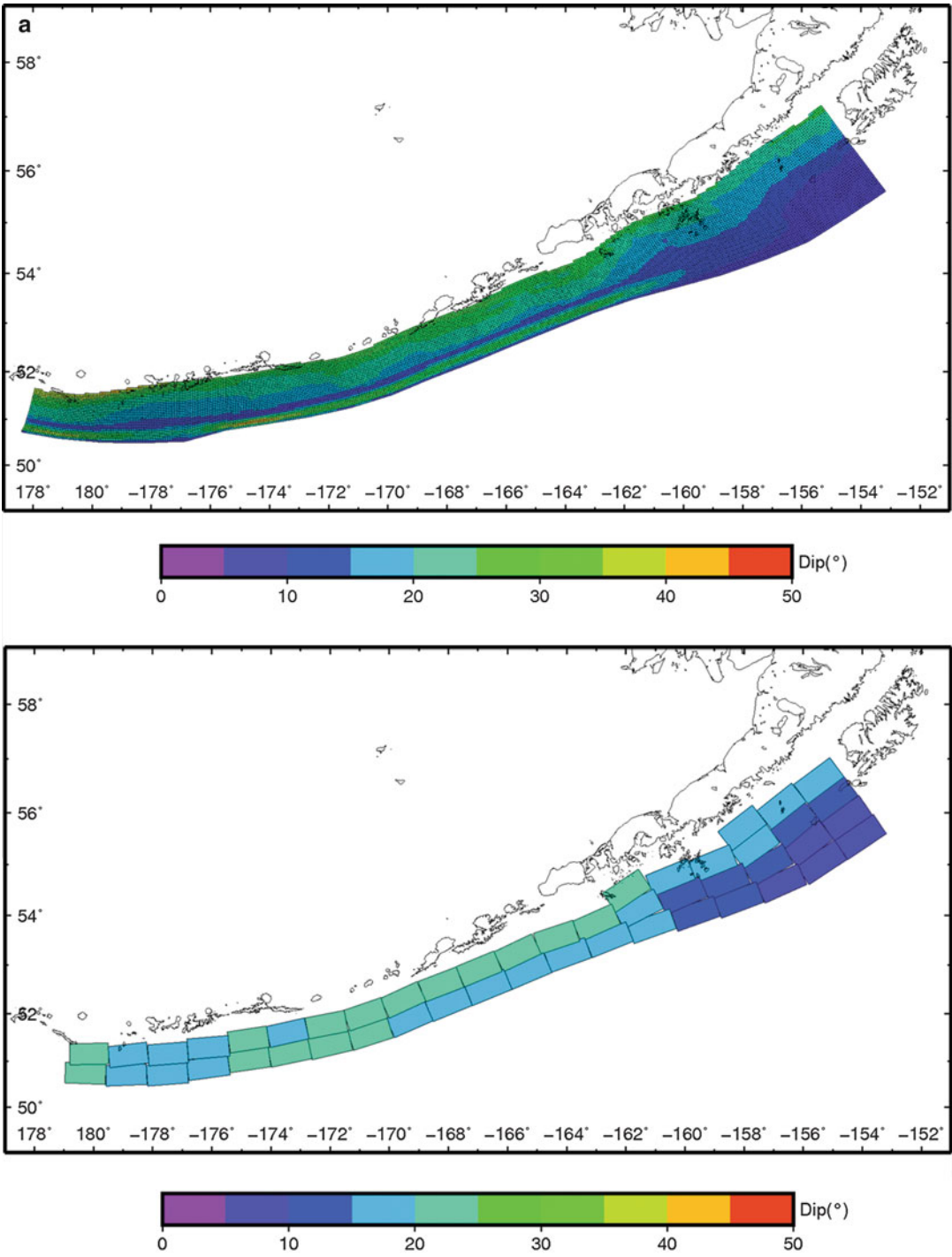
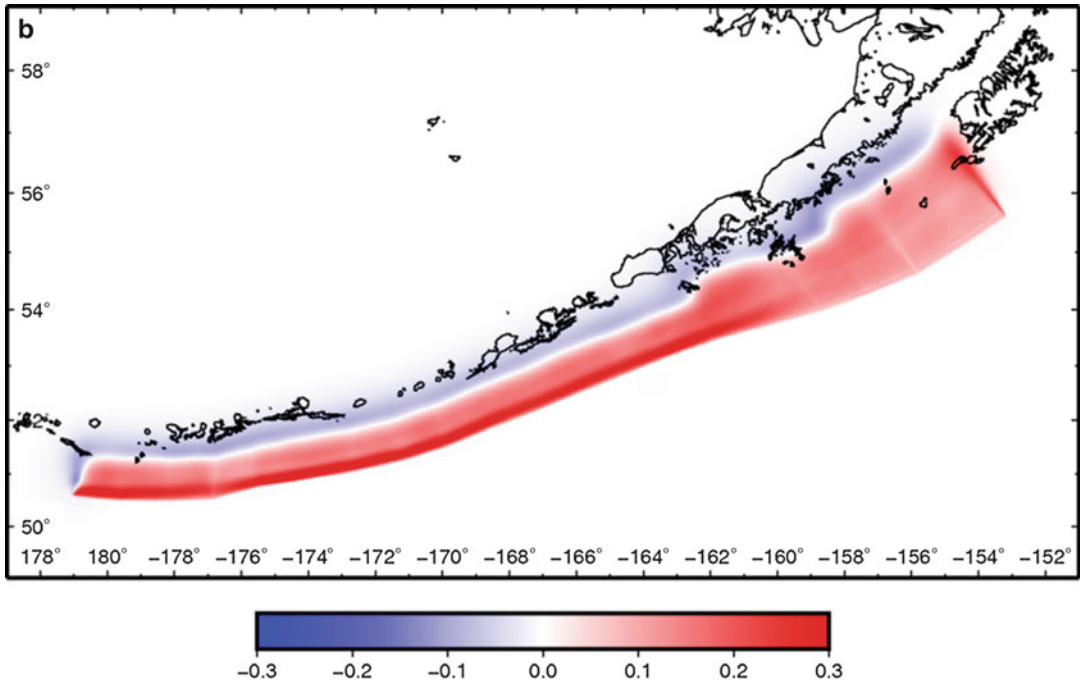


Fig. 2 (continued)



Tsunami Hazard and Risk Assessment on the Global Scale, Fig. 2 Illustration of development of source models for the Aleutians source zone showing (a) 5×5 km subfault model approximating the non-linear fault interface; (b) 100×50 km subfault model used as the

basis of the probabilistic calculations; and (c) aggregate surface co-seismic deformation for 1 m of slip on all component 5×5 km subfaults of each large 100×50 km subfault

- Differences in calculated deformation between the models are small.
- It gives greater deformation, and is therefore a more conservative choice, than island arc and forearc classes.
- The spatial resolution of the Crust 1.0 model is probably too coarse to accurately resolve spatial variations in crustal structure along subduction zone geometries (Laske pers. comm. 2013), justifying use of a spatially uniform crustal structure.

Recurrence parameters for global subduction zones have been compiled in the GEM subduction zone database (Berryman et al. 2013). Due to the computational and data management challenges of

developing a full probabilistic suite of tsunami events at the global scale, the full range of epistemic uncertainty in recurrence properties is not sampled. Instead, uncertainty is partially sampled through the use of a logic tree with the following approach:

- A truncated Gutenberg-Richter (exponential) magnitude-frequency distribution is considered. For earthquakes between a given minimum (m_0) and maximum magnitude ($m_{\max}$), this distribution describes the mean annual rate of earthquakes λ_m as (Kramer 1996)

$$\lambda_m = e^{(\alpha - \beta m)} \left(\frac{e^{-\beta(m-m_0)} - e^{-\beta(m_{\max}-m_0)}}{1 - e^{-\beta(m_{\max}-m_0)}} \right)$$

$$m_0 \leq m \leq m_{\max}$$

Tsunami Hazard and Risk Assessment on the Global Scale, Table 1 Crustal properties used for deformation calculations

Depth (km)	P-wave velocity (km/s)	S-wave velocity (km/s)	Density (kg/m ³)	Shear modulus (GPa)
0.00	2.50	1.20	2.10	3
1.00	4.00	2.10	2.40	11
15.00	6.00	3.50	2.72	33
20.00	6.50	3.74	2.82	39
25.00	7.10	4.04	2.99	49
35.00	8.00	4.50	3.30	67

- where α is the annual number of earthquakes above m_0 and β describes the number of large earthquakes relative to the number of small earthquakes. $\beta = 2.303b$, where b is the Gutenberg-Richter b -value from the unbounded definition $\lambda_m = 10^{(a-bm)}$. This model assumes a Poisson process where earthquakes are independent of each other.
- The *preferred* maximum magnitude ($M_{\max}$) from the GEM database has a weight of 0.7 and the *maximum* $M_{\max}$ a weight of 0.3. For sources that aren't in the GEM database the reported $M_{\max}$ is given a weight of 0.3, and a weight of 0.7 to $M_{\max}-0.1$ magnitude units is arbitrarily assigned.
- Full coupling is assumed everywhere.
- The preferred Gutenberg-Richter b -value from the GEM database is used. The observational record of earthquakes is used to calculate the b -value.
- The maximum slip-rate as defined below.

A uniform slip-rate for each source region is applied even though these vary along strike. A conservative approach is taken by using the maximum value reported along the full length of strike. In most cases differences are not large, although for the Ryuku, Kermadec, Tonga, and New Hebrides trenches there is a greater than 50 mm/year variation in slip-rate along strike from one end of the fault to the other (Berryman et al. 2013). Therefore, this assumption may lead to an overestimation of hazard for some areas. Rake is assumed to be perpendicular to strike (i.e., pure thrust motion) for subduction sources. Again this assumption is broadly conservative, as

there exist some sources where slip is expected to be much more oblique, such as the Puerto Rico Trench (ten Brink 2005).

Tsunami Propagation

Linear shallow water wave theory is applied for the offshore tsunami propagation. This means that in sufficiently deep water, the tsunami height can be constructed by the summation of the responses from multiple subfault sources. Hence, tsunami propagation simulations are only carried out once for each subfault, for unitary slip, and the hazard for each event is then determined by superposition of tsunami wave forms for the subfaults with scaled slip that construct that event. Here, a linear finite difference model (Satake 1995) is used for simulation of tsunami propagation for the simulation of the unit subfault sources on a 1 arc-min grid sampled from the GEBCO-08 global digital elevation model. In the linear simulations, tsunami waveforms are calculated at so-called hazard points. The hazard points are located sufficiently deep to utilize the linearity; that is, they are placed approximately along the 100 m depth contour at a spacing of 25 km, subject to constraints that they are not too near or too far away from the coastline. As the hazard points are normally not situated at exactly 100 m depth, the surface elevation is normalized to 100 m using the standard Green's law for shoaling of shallow water waves. For each event in the synthetic earthquake catalogue, the surface elevation is calculated at each hazard point near the coast by summing the contributions from the scaled subfaults that make up that event. For each site, this results in a list of

tsunami events with a maximum surface elevation and representative wave period, and the associated annual probability of the event.

Amplification of Offshore Surface Elevation to Maximum Run-Up Height

The standard way of estimating the tsunami run-up and inundation maps is to apply depth averaged NLSW models that apply drying-wetting schemes. Reviews of various methods used for simulating inundation, including the hydrodynamically more elaborate nonhydrostatic methods, are given by, e.g., Synolakis et al. (2007), Pedersen (2008), Lynett, ▶ [“Tsunami Inundation, Modeling of”](#) entry, and Løvholt et al. (2013). However, to estimate tsunami run-up height globally, these detailed numerical inundation simulations are too computationally intensive. A faster procedure is to relate the near-shore surface elevations at the hazard points to the maximum shoreline water levels by using a set of amplification factors based on the characteristics of the incident wave and the bathymetric slope. The surface elevation at the shoreline then acts as an approximation for the maximum inundation height or run-up height along the shoreline. The amplification factor procedure was developed by Løvholt et al. (2012a), and only a part of the procedure is reviewed here.

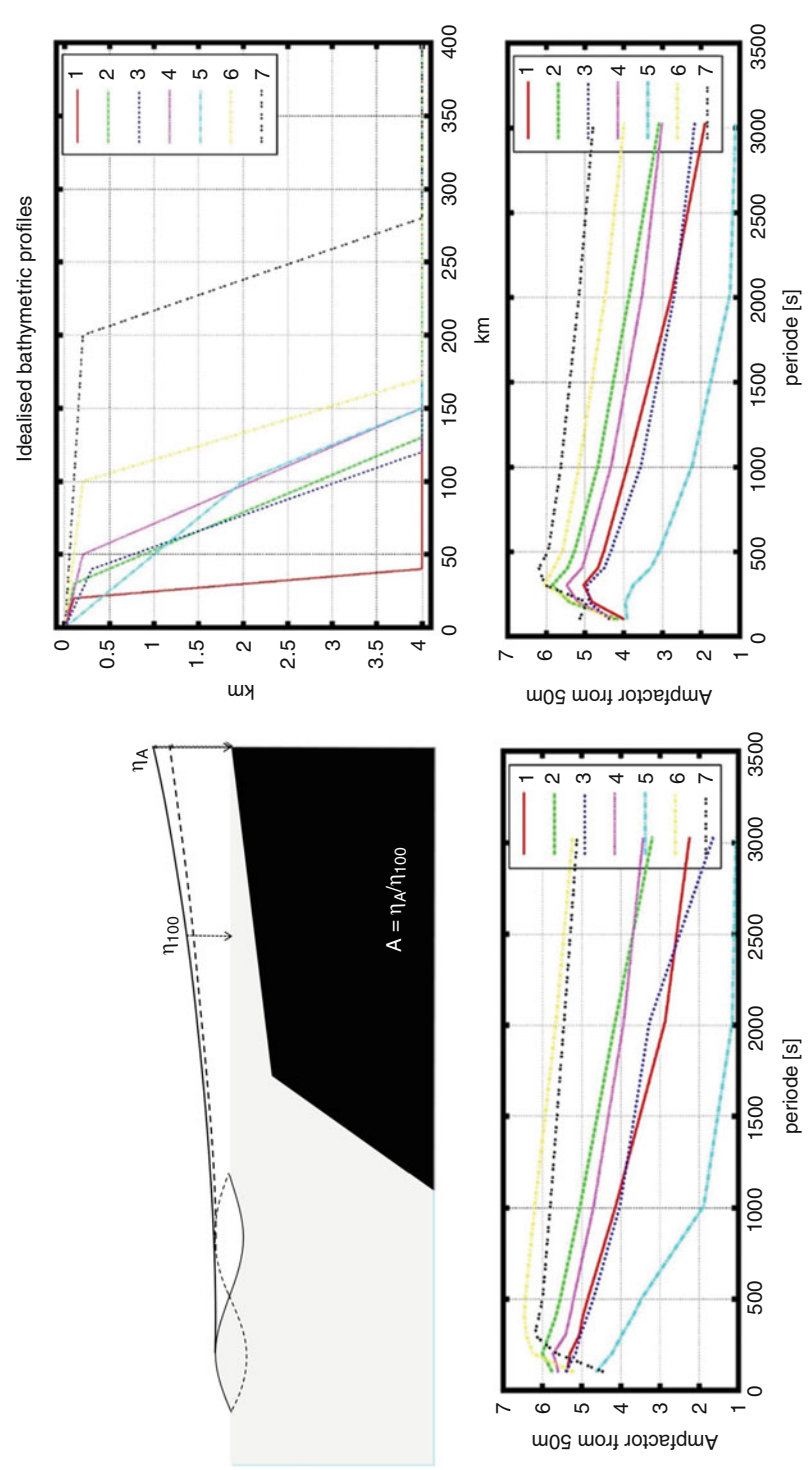
The amplification for a range of harmonic waves with different polarity and wave periods from a depth of 100 m to the shoreline is considered. The plane wave simulations are all run on idealized plane bathymetric configurations where the shelf is broken up into two linear segments. From the plane wave simulations, factors for amplification that relate the surface elevation at time series gauges located at water depths of 100 m to the maximum shoreline water level are computed and stored in lookup tables.

To determine the amplification factors along the idealized bathymetric profiles we apply a linear hydrostatic plane wave model. For smaller islands the plane wave assumption is severely violated, and a wave propagation model in two horizontal dimensions (2HD) is employed (Pedersen and Løvholt 2008; Løvholt et al. 2008). In the 2HD simulations, the spatial

and uniform resolution was 1 arc min (about 1.85 km). The models apply a no-flux boundary condition at the shoreline leading to a doubling of surface elevation due to reflection. Although the models do not include dry land inundation, the surface elevation on the boundary close to the shoreline (at 0.5 m water depth) with a no-flux condition yields a good approximation. For long nonbreaking waves, the linear solution for the water level at the shoreline and the nonlinear solution for the run-up height on land are identical (Carrier and Greenspan 1958). Based on Pedersen (2011), we may further assume that the procedure should also provide reasonably accurate results for waves of moderately oblique incidence.

Figure 3 illustrates the amplification of the incoming wave as a function of the water depth. Waves with either a leading trough or a leading depression are considered. As expected, the shorter waves amplify the most, and a leading depression gives more amplification, as described by Tadepalli and Synolakis (1996). We found that the amplification factor from a depth of 50 m ranges from nearly unity (steep slope combined with wide earthquake or earthquake located close to or partly on land) up to about 6 (very gentle slopes, e.g., profile 7 in Fig. 3). In the first case, the wave is much longer than the distance from the hazard point to the shoreline, and hence both locations are influenced by the reflection at the shoreline at the same time. However, for shorter waves (i.e., short compared to the distance between the shoreline and the time series gauge) the amplification follows Green's law until the waves reach the shoreline and are doubled in height due to the reflection here (the reflection is not taken into account in Green's law). On the other hand, the amplification factors for small islands (with a diameter of the same order or less than the wavelength) with the same set of parameters are in the range 1–3. In the plane wave and 2HD model, the estimated maximum shoreline water elevation is measured close to the shoreline (at 0.5 m water depth).

To assign an amplification factor, we need to know the bathymetry profile near each of the amplification factors. However, a general method



Tsunami Hazard and Risk Assessment on the Global Scale,
Fig. 3 Amplification factors of the surface elevation from a water depth of 50 m and to the shore as a function of the incident wave period. Additional amplification from 100 m water depth is accounted by Green's law (an additional amplification factor of $2^{1/4}$). *Upper panel*, idealized trough. *Mid panel*, amplification factors for a harmonic wave with leading trough. *Lower panel*, amplification factors for a harmonic wave with leading elevation

Tsunami Hazard and Risk Assessment on the Global Scale, Table 2 Comparisons of run-up calculations using amplification factors and the standard run-up

model MOST. Comparisons are made for two locations along the Indian shoreline, namely, Chennai and Nagapattinam

Source location	Mw	Runup Chennai		Runup Nagipattinam	
		MOST	Amp factor	MOST	Amp factor
Andaman	8.25	0.4	0.4	0.5	0.4
Andaman	8.75	1.8	1.7	1.9	2.0
Andaman	9.25	5.6	6.2	5.6	7.6
Sumatra	8.45	–	<0.3	0.2	<0.3
Sumatra	9.45	–	1.2	1.4	1.6

for obtaining this automatically for any point along the coastline is not yet in place. Instead, we assign each idealized bathymetric profile manually to a section comprising a range of points with relatively similar cross-sectional depth profiles. To estimate the maximum shoreline water elevation from the offshore time series gauges in a tsunami simulation, the amplification factor for a set of parameters is extracted from the lookup tables and in turn multiplied with the maximum surface elevation measured at the hazard points.

The first comparison of the procedure towards more accurate models was presented by Løvholt et al. (2012a). As an example, we present here a new comparison in Table 2, comparing the simulated maximum inundation using the NSW MOST model (Titov and Gonzalez 1997) with the one obtained using amplification factors for the two Indian cities of Chennai and Nagapattinam. The comparison was conducted using five of the different scenarios originating from the Sunda Arc (Andaman and Sumatra trenches).

Tsunami Inundation and Exposed Assets

In the previous subsection, we presented an approximate method for estimating the maximum run-up in a proximal area at any location along the shoreline. In the absence of detailed inundation models, the run-up height at the shoreline obtained from methods such as the amplification factor method needs to be extrapolated onshore if it is to be used in further hazard mapping and risk analysis. In both PTHA and PTRa, this would

also need to include the uncertainties in the tsunami physical intensities to be extrapolated. To this end, we use a simple extrapolation technique to derive inundation maps and flow depth probabilities. This method takes the maximum water level at the calculation points along the coastline contour and then assumes that locally, the onshore water level is uniformly elevated above the mean sea level. Then, an inverse distance weighted method is used to extrapolate the maximum water levels at the shoreline to the inland DEM, using the SRTM elevation dataset.

In locations where the topography near shore is extremely flat, the inundation zone could extend an unrealistic distance inland, and the inundation in principle would be unbounded for an infinitely low slope. Using standard inundation tools, the horizontal inundation is limited by the flow resistance, usually formulated through a quadratic resistance term such as the Manning friction or similar. For the global analysis, a simpler approach needs to be applied in order to take into account the pressure head loss due to friction during overland flow; therefore, it was assumed that the tsunami is subject to a constant pressure head loss due to quadratic friction along the path normal to the shore toward inland. In this way, a maximum limit for the horizontal inundation may be computed.

In the analysis presented herein, we represent both source uncertainties (such as fault geometry and variable slip) and uncertainties due to modeling uncertainties as an integral standard deviation σ . We label this as the aleatory uncertainty and define it as a quantifiable uncertainty that can be

represented by data. This uncertainty is distinguished from the epistemic uncertainty that accounts for our lack of knowledge. While the borderline between the aleatory and epistemic uncertainties is not always clear, a practical consideration for the epistemic uncertainty is that it accounts for the uncertainty between models, quantifying it among different experts' opinion.

For the representation of the hazard, we use a lognormal probability distribution of the overland flow depth in our further analysis. For a given (dry-land) cell, we may then define the following lognormal probability density function for the flow depth H :

$$P(H) = \frac{1}{\sigma_H \cdot H \cdot \sqrt{2\pi}} e^{-\frac{(\ln(H) - \ln(\bar{H}))^2}{2\sigma_H^2}}$$

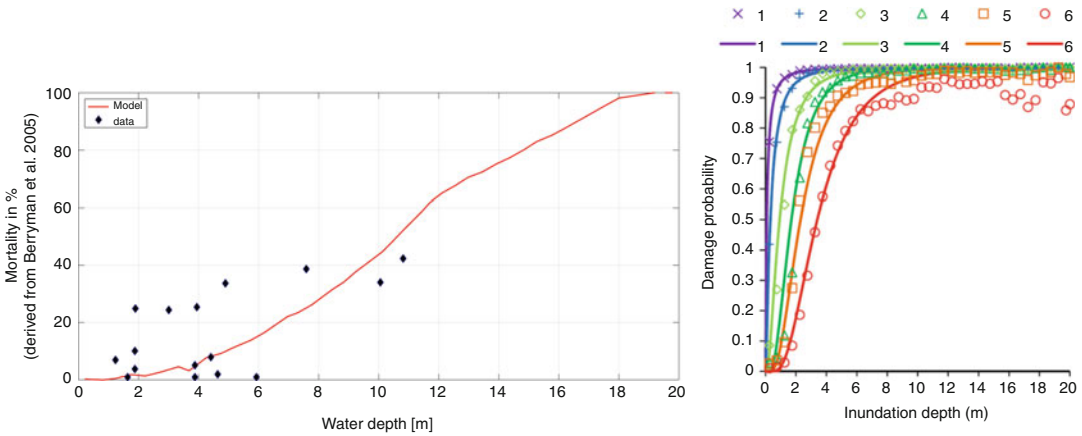
Here, the median flow depth $\bar{H}$ in a given cell is taken as $R \cdot D$, R being the simulated run-up for a given event (as, for instance, derived from the amplification factor method) and D being the topographic altitude. The reason for choosing the flow depth as the tsunami physical intensity is that the vulnerability functions, which are used to estimate the economic losses later in this assessment, are solely based on H . Due to practical reasons, as indicated above, we label the uncertainty as aleatory, although it is clearly also related to our lack of knowledge. A logarithmic uncertainty $\sigma_R = 0.35$ for the run-up height R is adopted, largely based on experience from PTHA conducted elsewhere (see, e.g., Thio et al. 2010 for a discussion). Using this representation, the uncertainty related to the flow depth is expected to vary with flow depth and needs to be normalized. A small flow depth will then be subjected to a larger uncertainty than a large flow depth. In the present probabilistic analysis, we employ a variable flow depth uncertainty given by $\sigma_H = \sigma_R(1 + \log[R/\bar{H}])$. Generally, both the flow depth and the inundated area should be subjected to a level of uncertainty. In the present examples, however, the inundated area is fixed for a given tsunami scenario, while the flow depth H , as described above, has an uncertainty level associated. The

method explained above is used to derive inundation maps using the STRM 90 m spatial resolution dataset, providing maps of the hazard intensities that may be convolved with the physical vulnerability (see below).

Building Stock Vulnerability, Mortality, and Exposed Population

Vulnerability has different dimensions (e.g., physical damage, economic loss, loss of life, social loss), and in this global analysis only the physical one is considered. This is done by means of vulnerability functions, which are a continuous representation of the damage functions. Each vulnerability function relates a tsunami physical intensity to an expected mean damage ratio and an uncertainty measure. At present, vulnerability usually refers either to human mortality ratio (probability of fatality) or building susceptibility to damage. In both cases, the vulnerability associated to tsunami is derived from historical tsunami damage databases. While empirical data were limited prior to the string of catastrophic events between 2004 and 2011, systematic use of data from post-tsunami field damage surveys, in particular from the 2011 Tohoku tsunami, have provided a more comprehensive dataset and allowed the development of fragility curves, useful to derive vulnerability functions. An example of two such datasets, one for tsunami mortality and one for building fragility, are shown in Fig. 4. It is noted that the frequent use of flow depth stems from its direct availability from field data. However, other physical intensities such as the current velocity and momentum flux, which are less easily measured, may be used to estimate damages and losses from a theoretical standpoint, since they are more directly linked to the tsunami load function. A review of various models for establishing the physical vulnerability can be found in Gonzalez et al. (2009), Maqsood et al. (2014), and Løvholt et al. (2014b).

The use of vulnerability functions allows consideration of the uncertainties in the structural behavior of buildings when subjected to a given tsunami intensity. For each building class, a unique vulnerability function is defined by the expected mean damage ratio (i.e., the repair-to-



Tsunami Hazard and Risk Assessment on the Global Scale, Fig. 4 *Left:* tsunami mortality derived from post tsunami studies (Berryman et al. 2005). *Right:* tsunami

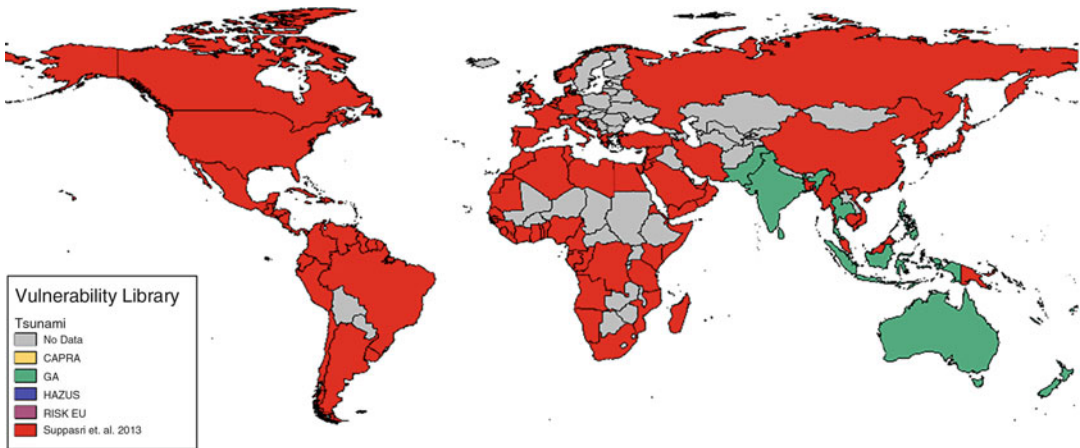
building fragility curves from measured flow depth following the 2011 Tohoku earthquake and tsunami (Suppasri et al. 2013)

replacement cost relative to the total value of the element) associated to each intensity level and by the uncertainty.

While the above examples comprise cases of tsunami damage functions that may be used to quantify risk, building vulnerability may not be immediately conveyed to any building worldwide but must depend on the design, materials, and construction of the building. Similarly, the mortality ratio is expected to be a multifaceted function of many factors in addition to the flow depth, such as tsunami warning time, season, time of day, population distribution, population age, population tsunami awareness, and building design among others. The mortality ratio is expected to vary by more factors and be more heterogeneous than the building fragility and associated damages and losses. Hence, the economic risk due to building damage is expected to be more easily established than the mortality risk. Because of that, the direct economic loss was chosen as the risk indicator for the GAR15 report (UN-ISDR 2015). To work out a sufficiently comprehensive dataset taking into account the global variability of datasets and including the regional knowledge in the global model, a workshop was held with the purpose to develop vulnerability functions based on expert opinions for a wide range of buildings (Maqsood et al. 2014). In summary, for countries located

within the Asia-Pacific region, the vulnerability functions developed under the framework of the experts' workshop were used, considering the local characteristics of the built stock and the particularities of them. For the rest of the world, vulnerability functions from the fragility curves by Suppasri et al. (2013) derived from the 2011 Tohoku tsunami field data were employed. The procedure to convert the fragility curves into vulnerability functions was the one proposed by CIMNE et al. (2013). Figure 5 shows the tsunami vulnerability regionalization implemented for the global tsunami risk assessment where the ones denoted as GA correspond to the functions generated during the experts' workshop while the functions denoted as Suppasri et al. (2013) were used for the other regions.

As an example, Fig. 6 shows two different vulnerability functions for unreinforced masonry residential buildings. Figure 6a shows the vulnerability function derived from the Suppasri et al. (2013) fragility curves, while Fig. 6b shows the vulnerability function generated from experts' opinion for the Asia-Pacific region. For both cases both the expected loss and its variance (σ^2) are shown. From the figures the differences in the functions are clear, and therefore, the importance of considering regional variations, even when assessing risk at global level, becomes evident. The complete set of tsunami



Tsunami Hazard and Risk Assessment on the Global Scale, Fig. 5 Tsunami vulnerability functions regionalization (CIMNE and INGENIAR 2015)

vulnerability functions used in this global PTRAs can be found in CIMNE and INGENIAR (2015).

Probabilistic Loss Estimation for the Built Stock

PTRA provides a robust probabilistic framework, and considers probability of losses due to synthetic events. Risk is expressed in terms of exceedance rates (or probabilities within a given time period) which specify the frequencies with which hazardous events cause losses equal or higher than specified values. This facilitates a comprehensive disaster risk management scheme as outlined by Cardona (2009) which, as outlined earlier, is important for tsunamis where historical records may not be sufficiently complete (see, e.g., Cardona et al. 2014).

The global probabilistic multirisk assessment for GAR15 (UN-ISDR 2015), compiled from hazard-specific risk assessments such as PTRAs, was performed using the CAPRA¹ Platform (Cardona et al. 2012), an open-source set of hazard, exposure, vulnerability, and risk modules which incorporates state-of-the-art methodologies developed by the Consortium Evaluación de Riesgos Naturales – América Latina (ERN-AL).

The CAPRA Platform allows multihazard risk assessments at different resolution levels always applying the same probabilistic arithmetic to calculate the risk metrics that are discussed below.

Since risk is quantified in terms of loss exceedance rates, a probabilistic representation is needed for both hazard and vulnerability. For the hazard, the temporal probability needs to be considered besides the geographic distribution. To perform a fully probabilistic risk assessment, an event-based approach is required (Salgado-Gálvez et al. 2014). The hazard therefore is represented by means of a set of stochastic events whose ideal requirements are

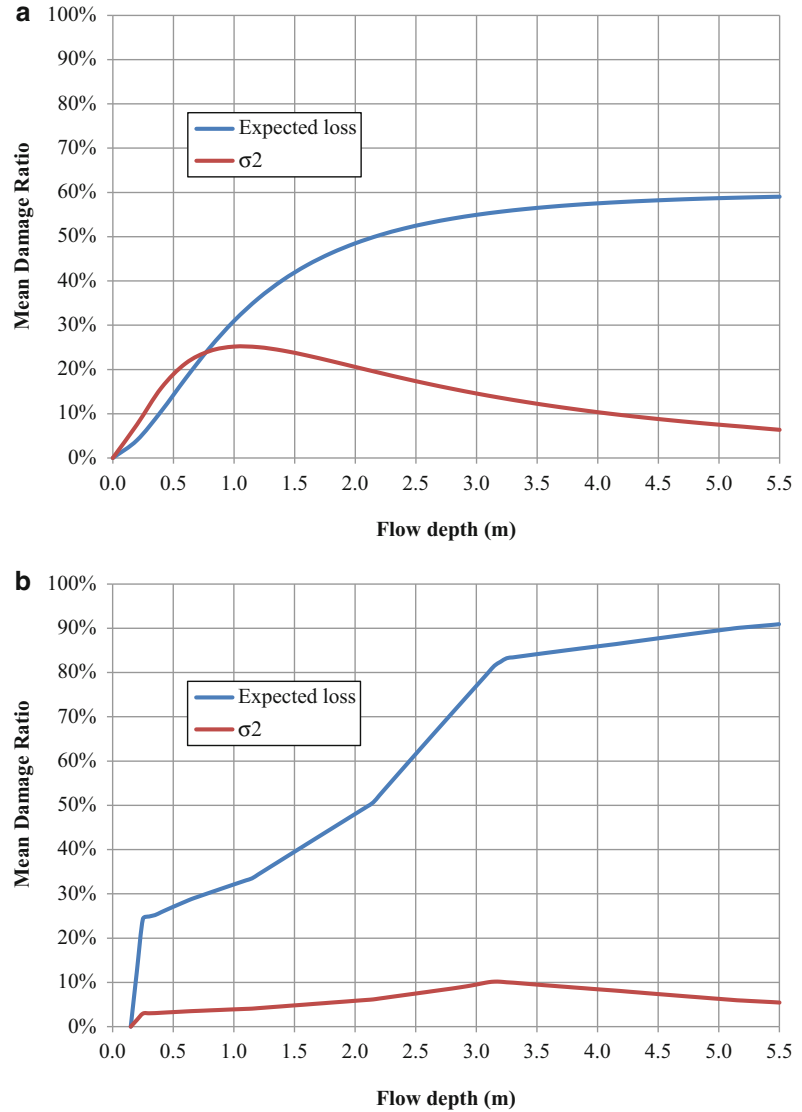
- To be mutually exclusive.
- To be collectively exhaustive.
- To allow a probabilistic representation.
- To have assigned, each of them, an occurrence frequency.

The probabilistic representation means that the hazard intensities are represented at least by the first two moments (related to the mean and standard deviation) of the (lognormal) PDF. In this case, the PTHA is represented by means of a synthetic set of tsunami events that complies with the requirements listed above and hence suitable for the subsequent fully probabilistic damage and loss estimation.

¹Comprehensive Approach to Probabilistic Risk Assessment (www.ecapra.org)

Tsunami Hazard and Risk Assessment on the Global Scale, Fig. 6

Tsunami vulnerability functions for unreinforced masonry buildings. (a) based on fragility curves. (b) based on expert opinion (CIMNE and INGENIAR 2015; Maqsood et al. 2014)



According to the analytical procedure proposed by Ordaz (2000) and used in the CAPRA platform, the probability density function for the loss on the j^{th} exposed asset, conditional to the occurrence of the i^{th} scenario, is defined as $f(l_j|Event_i)$. Because it is not possible to compute this probability distribution directly, it is computed by chaining two separate conditional probability distributions where the first part has to do with the vulnerability (the expected loss given a hazard intensity) and the second with the hazard

(the hazard intensity given the occurrence of the event):

$$f(l_j|Event_i) = \int_0^{\infty} f(l_j|Hi) \cdot f(Hi|Event_i) dHi$$

where Hi is the hazard intensity. Also, the probability density function of the loss for each scenario is computed by aggregating losses from each asset included in the exposure database

(De Bono and Chatenoux 2014). A set of stochastic events, independent of time, is generated. Then, for each scenario and for each asset included in the exposure database, considering its geographic location and the estimation of the hazard intensity at that point, the expected loss and its variance are calculated using the associated vulnerability function. This is repeated for all the assets included in the exposure database to yield the loss PDF for each event. When this procedure is concluded for all the events in the stochastic set, the loss exceedance rate for the whole portfolio is calculated from the event's loss probability distribution functions and their occurrence frequencies.

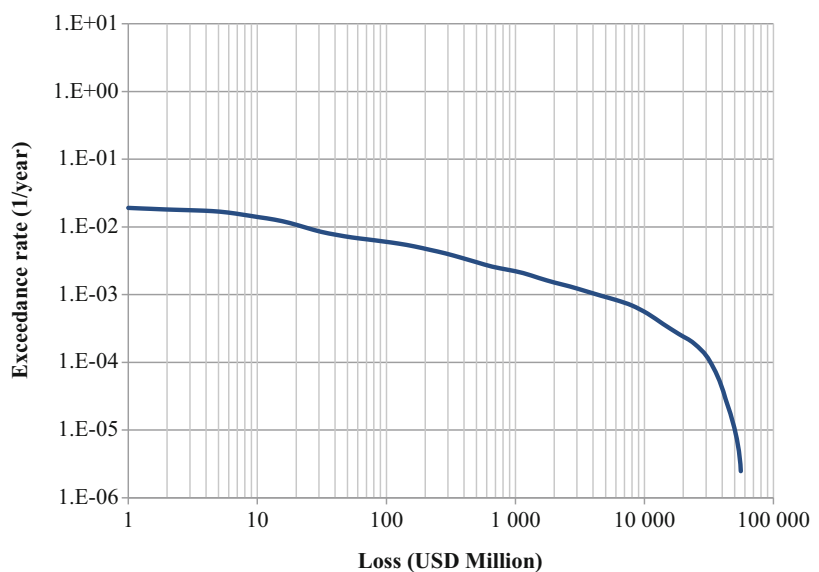
The main output of a fully probabilistic risk assessment is the loss exceedance curve (LEC) which relates different loss values with their corresponding exceedance rates, usually expressed in annual terms. LEC is calculated using the following equation and can only be obtained if an event representation is available for the hazard (in contrast to, e.g., a hazard curve that is decoupled from event information)

$$v(l) = \sum_{i=1}^N \Pr(L > l | \text{Event}_i) \cdot F_A(\text{Event}_i)$$

where $v(l)$ is the rate of exceedance of a loss l , N is the total number of hazard events, $F_A(\text{Event}_i)$ is the annual frequency of occurrence of the i^{th} hazard event, while $\Pr(L > l | \text{Event}_i)$ is the probability of exceeding the loss l , given that the i^{th} event occurred. The sum in the equation is made for all potentially damaging events. When calculating $\Pr(L > l | \text{Event}_i)$ using this methodology, the probabilistic representation of the hazard plays a relevant role since both probability moments are combined with the mean and uncertainty values of the vulnerability functions (Miranda 1999; Ordaz 2000; Ordaz et al. 2000) to obtain the subsequent damage and loss levels in a consistent probabilistic manner. Figure 7 shows a schematic representation of a tsunami LEC from where some of its features can be seen, most notably the approximately exponential tail of the distribution describing the much greater frequency of small losses compared with large losses.

Although the LEC contains all the relevant risk information, for comparison and communication purposes it is sometimes preferred to use metrics that are related to a specific value such as the average annual loss (AAL) and the probable maximum loss (PML) for a mean return period of interest. The first can be computed as

Tsunami Hazard and Risk Assessment on the Global Scale,
Fig. 7 Schematic tsunami LEC



the area under the LEC and represents the expected annual loss, over a sufficiently long time frame considering all events. It is a robust risk metric useful for comparison purposes which is not very sensitive to the uncertainties (Marulanda 2013). The inverse value of the exceedance rate corresponds to the mean return period; this value represents the average time for a loss value to be observed and can be directly read from the LEC. Although there is not a standard to define the return periods of interest, this risk metric is usually associated with low occurrence events and hazards, such as the tsunami, where return periods between 500 and 2,500 years are of interest. At this stage it is important to highlight that because there is correlation in the losses (since they are being caused by the same events), the return period of the loss does not necessarily correspond to the return period of the hazardous event (Salgado-Gálvez et al. 2015a). Another common representation of the risk results are the PML plots where the same information included in the LEC is rearranged. Instead of exceedance rates the temporal variation is represented through mean return periods, Fig. 8 shows a schematic tsunami PML plot.

A good example to better understand the concept of mean return periods is by

calculating the probability of having a loss equal to a T year return period in the next T years. That can be calculated by means of the following equation:

$$Pe(l, T) = 1 - e^{-v(l)T}$$

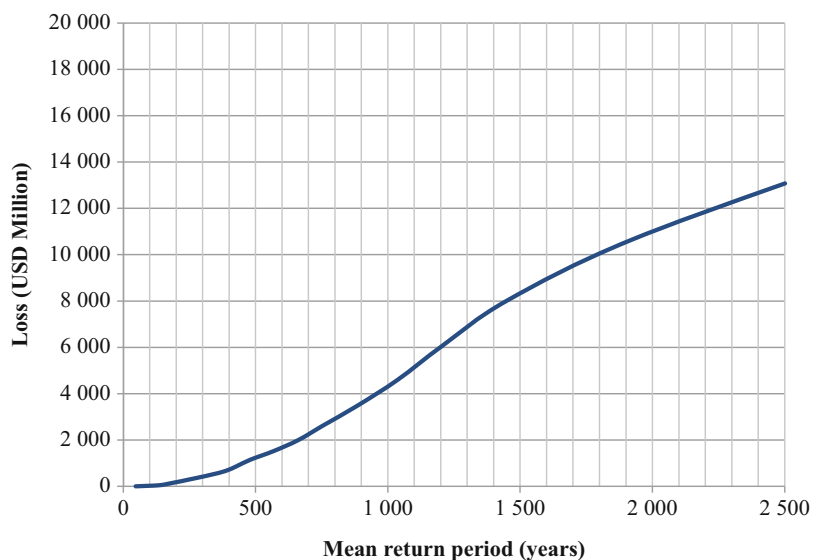
where $Pe(l, T)$ is the probability that the loss l is exceeded in the next T years and $v(l)$ is again the time-independent loss exceedance rate.

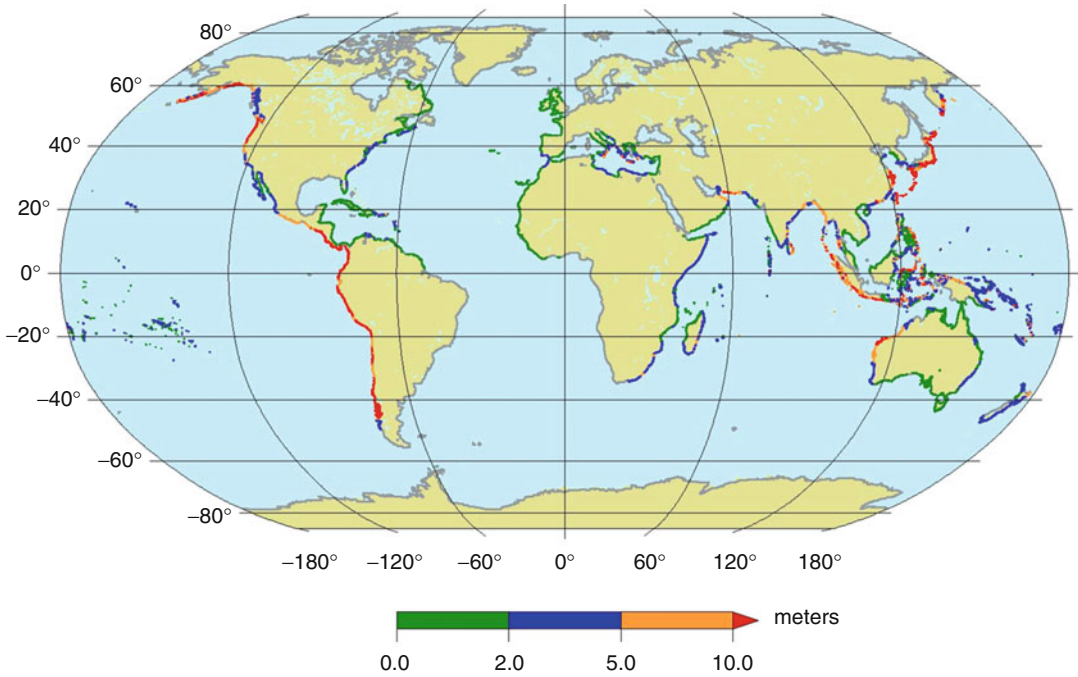
Global Tsunami Hazard and Loss Estimation

Tsunami Hazard and Exposure

Using the methodologies presented above, we may, for fixed return periods, derive tsunami hazard maps from the stochastic event set and estimate the number of exposed elements at risk. However, the first full global tsunami hazard analysis by Løvholt et al. (2014a) used the scenario-based method. The underlying assumption in their analysis was similar to the one used in the probabilistic method employed for the fully probabilistic risk assessment below, namely, that the tectonic slip rates are used to determine earthquake slip over a 500 year mean return period, assuming that the subduction zone faults are

Tsunami Hazard and Risk Assessment on the Global Scale,
Fig. 8 Schematic tsunami PML plot





Tsunami Hazard and Risk Assessment on the Global Scale, Fig. 9 The global tsunami hazard map displaying the run-up height for a mean return period of 500 years

locked. Figure 9 shows the global tsunami hazard map of Løvholt et al. (2014a), depicting the run-up height projected along different shorelines. Areas with red color have run-up heights exceeding 10 m, typically affecting Japan, western coastlines of South and parts of Central and North America, and Indonesia. Overall this analysis example predicts that areas adjacent to the Pacific Ring of Fire, the Sunda Arc in South East Asia, and the Makran trench south of Pakistan and Iran are subjected to the largest hazard levels. Other hazardous areas include the North Atlantic, the Mediterranean, and the Caribbean regions.

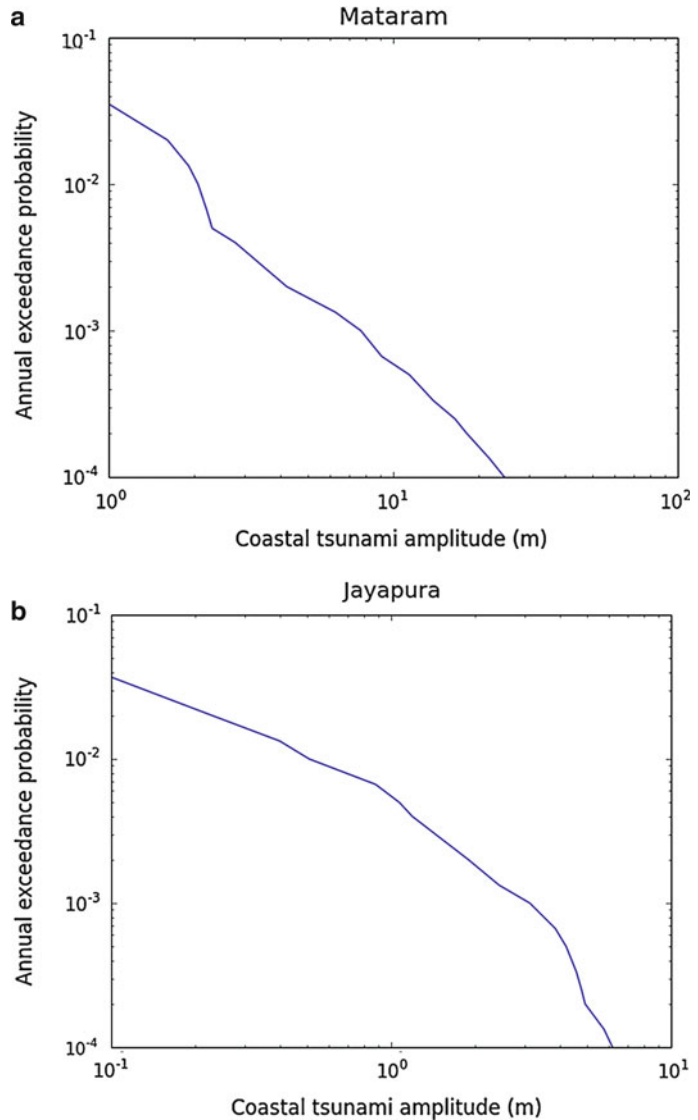
Using PTHA, every point on the deterministic hazard map depicted in Fig. 9 is represented by a hazard curve containing the annual probability of exceeding a certain tsunami height or run-up. Figure 10 shows example hazard curves for two locations in Indonesia from Horspool et al. (2014), representing the annual probability of exceeding a given coastal tsunami height. Tsunami hazard at a site can be disaggregated by source to determine which faults contribute

most to the tsunami hazard. Figure 11 shows disaggregation of 500 year return period hazard for the two locations shown in Fig. 10 (Horspool et al. 2014). For the city of Mataram, Lombok (Fig. 11a), we can see that although the Sunda Subduction Zone to the south is an important source of tsunami hazard, the Flores Back-Arc Thrust to the north contributes more to Mataram's tsunami hazard, even though maximum magnitudes on this source are lower than on the Sunda Subduction Zone. For Jayapura, Papua (Fig. 11b), the local North Papua Subduction Zone is the dominant source of hazard; however, tsunamis generated off the Philippines, Marianas, and Japan also contribute to the tsunami hazard. This information can help risk managers prioritize and plan for alternative scenarios.

An example of the predicted population exposed to the 500 year mean return period tsunami is presented in Fig. 12 (Løvholt et al. 2014a). The exposure was obtained by extrapolating the run-up height from Fig. 9 over the STRM topography, integrating the

Tsunami Hazard and Risk Assessment on the Global Scale,

Fig. 10 Example of a hazard curve (annual probability of exceedance as a function of coastal tsunami height) for (a) Mataram, Lombok and (b) Jayapura, Papua, from Horspool et al. (2014). See Fig. 11 for location. Note that horizontal scale is different between the two figures

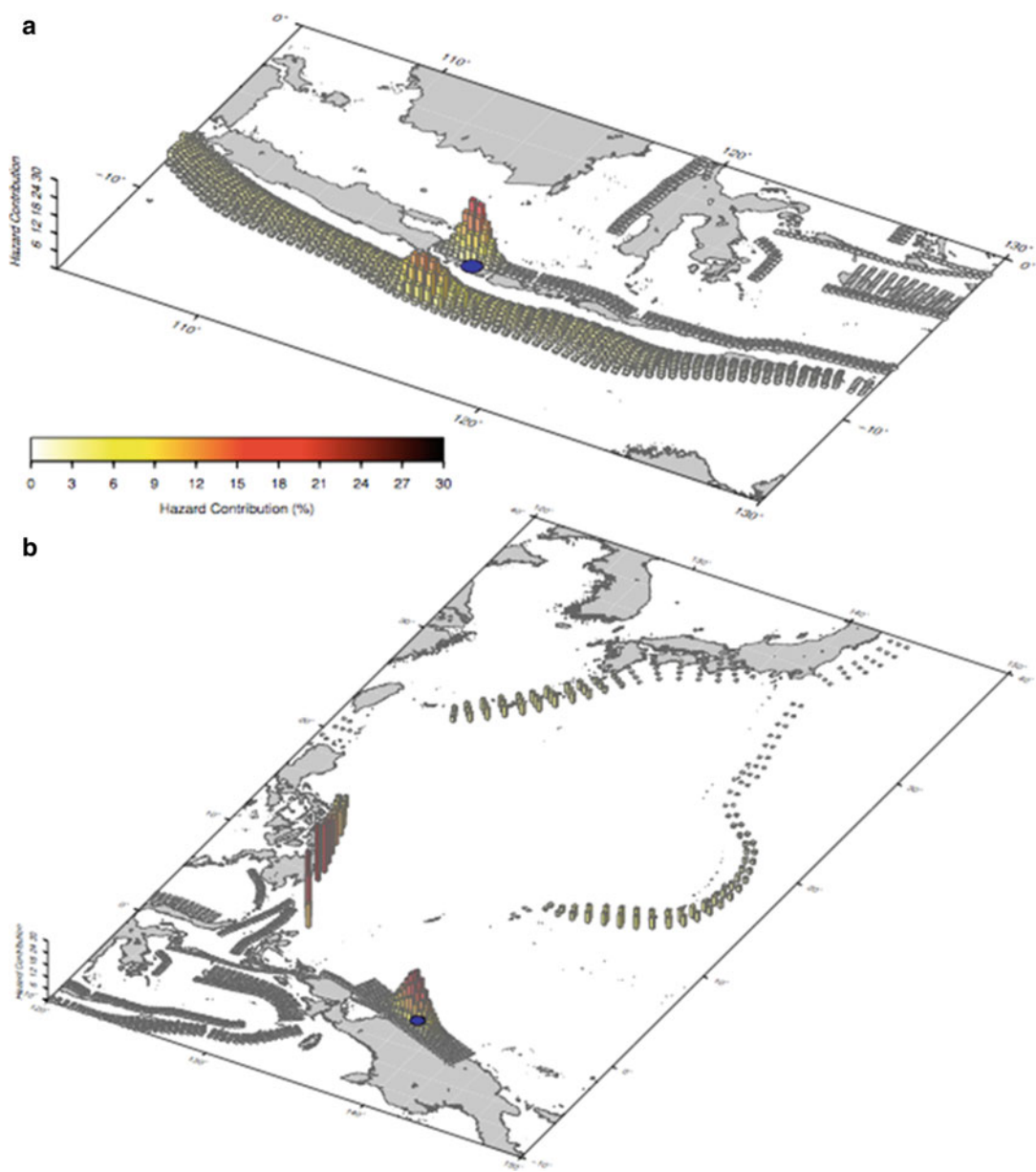


LandScan (2013) population data for each country. Figure 12 shows that populous Asian countries account for a large absolute number of people living in tsunami-prone areas (Fig. 12), with Japan standing out, accounting for about 50 % of the world's total according to this study (Løvholt et al. 2014a). With smaller coastal populations, countries such as Chile, Peru, and Ecuador have therefore a lower population exposure. Figure 12 further shows that smaller countries such as Macao and the Maldives have high population exposures when we normalize the exposure by the total population in each country.

At country level, the relative population exposure is considered a more relevant measure for the tsunami risk than the total exposure. Hence, the tsunami risk is as prominent for many smaller countries as Japan, despite the smaller absolute exposure. We will revisit this in our discussion related to tsunami risk below.

Tsunami Risk

Of the different dimensions of vulnerability for use in tsunami risk analysis, we have briefly discussed the physical building vulnerability and exposed population mortality. In this

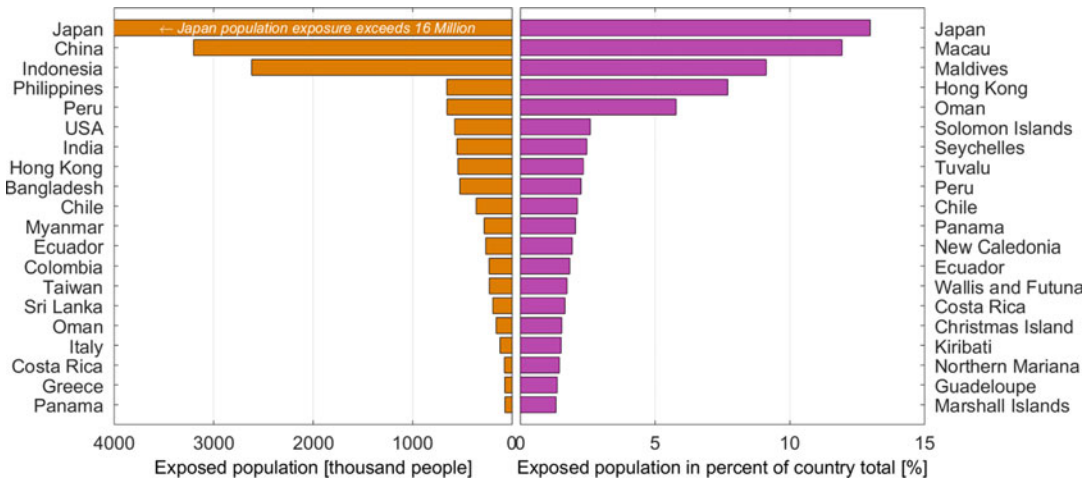


Tsunami Hazard and Risk Assessment on the Global Scale, Fig. 11 Example of hazard disaggregation for two locations in Indonesia, (a) Mataram, Lombok and (b)

Jayapura, Papua. Hazard disaggregation allows the relative contribution of different sources to be determined (Figure from Horspool et al. (2014))

subsection we will analyze the direct economic losses due to building damage as input to the economic loss metric. It is noted that the economic loss, in certain cases, may yield a very different risk than, for instance, using mortality data. We will not provide an in-depth discussion related to the different methods for evaluating the

tsunami risk, but note that the economic loss was chosen since it provides a simpler and less uncertain measure of the tsunami risk, besides enabling a more straightforward and direct comparison of the risk results with that associated to other natural hazards also considered in the global risk assessment (CIMNE and INGENIAR 2015;



Tsunami Hazard and Risk Assessment on the Global Scale, Fig. 12 Global tsunami population exposure for a 500 year return period (Løvholt et al. 2014a). *Left panel*;

number of people living in areas potentially affected by tsunamis. *Right panel*; number of exposed persons divided by the total population in each country in percent

UN-ISDR 2015). We add that index-based method with a more descriptive nature commonly applied for tsunamis (e.g., Dominey-Howes and Papathoma 2007) was not found suitable due to requirements for a fully probabilistic assessment to be undertaken.

For low-probability high-consequence events such as tsunamis, the annual average loss (AAL) will often yield very small losses since the potential losses are annualized over a very long exposure time frame. Therefore the AAL may be insufficient to fully understand the risk for natural hazards like tsunamis (CIMNE and INGENIAR 2015; Salgado-Gálvez et al. 2015b). Instead, the probable maximum loss (PML) for long mean return periods has been commonly adopted for quantifying tsunami risk (Dominey-Howes et al. 2010).

Figure 13 shows the absolute values of the probable maximum loss (PML) for three different return periods, 100, 500, and 1,500 years. In all cases, it can be seen that Japan has by far the largest PML values not only because of the high hazard level but because of its large and expensive exposure. Other countries subject to large PML values are countries with large coastal populations and infrastructure such as China, USA, Indonesia, the Philippines, India, and Australia. The relative trend is fairly consistent for all

the different mean return periods depicted herein. Naturally, we find increasing values for the PML with increasing return period. For Japan, we see a ten times increase from the 100 year to the 1,500 year return period, bearing in mind that it is not a general rule and the ratio depends on both hazard and vulnerability characteristics. For comparison purposes, it is better to normalize the expected loss by the total exposed value for each country, which provides a better picture of the risk on a country level. To this end, Fig. 14 shows the same PML values as in Fig. 13 but normalized by the total value of the built stock. For the 100 year return period, a limited number of countries appear on the maps and are mostly located in East and South East Asia or in the Pacific Coast of South America (the effect on Small Island States is not visible in these maps are discussed below). For the 500 year mean return period, more countries in South America, the eastern Mediterranean, and North America (USA) are visible. The results are explored up to 1,500 year mean return period, indicating that the tsunami risk for large return periods ranging from several hundreds to thousands of years are relevant for the majority of exposed countries.

Examples of PML curves for some selected countries are presented in Figs. 15 and 16 (Small Island States). Both the absolute PML and

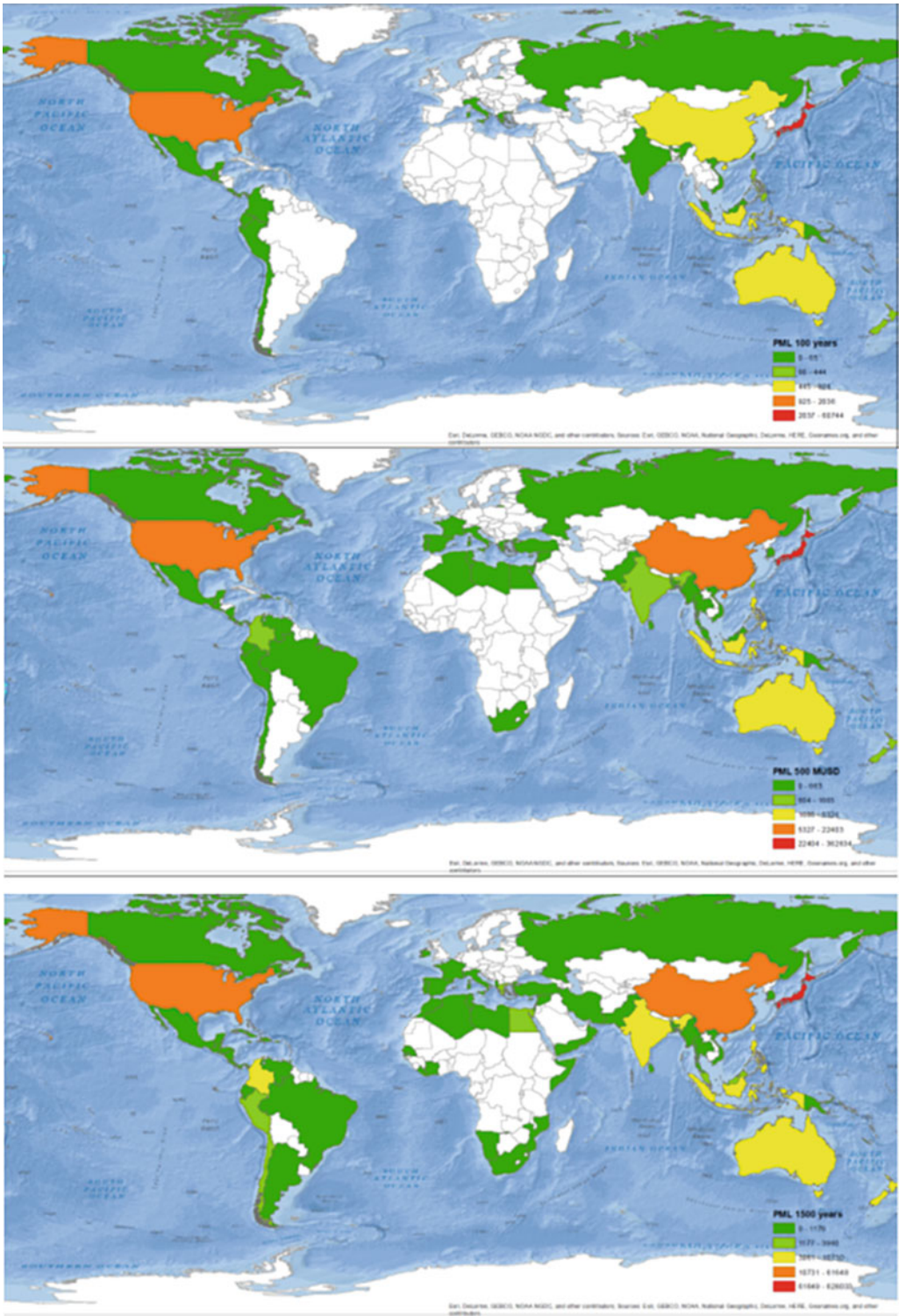


Fig. 13 (continued)

relative PML plots are shown. We note that missing data points in the figures indicate small losses. A typical trend is that the relative PML for Small Island States is often high. Here, we find that the largest relative PML values worldwide are found for Macao, Tonga, and Hong Kong, exceeding the relative PML values from highest ranked larger countries like Japan and the Philippines. It is also interesting to compare the slope of the PML curves. In Fig. 15, we see that while relative PML values for countries such as Egypt and Myanmar are low for the shorter return periods, they increase more rapidly than for other countries up to the highest return period considered as a consequence of the hazard rate (10,000 years). For the Small Island countries, a similar trend can be found for the Maldives. This emphasizes the dominance of the long return period tsunami risk, as noted above and which is the reason to apply fully probabilistic risk methodologies to quantify the potential losses.

The two most important tsunami events in recent times are the 2004 Indian Ocean tsunami (Stein and Okal 2005; Titov et al. 2005) and the 2011 Tohoku tsunami (Ozawa et al. 2011; Hooper et al. 2013). Based on the extensive losses originating from these events, some comparison from the probabilistic risk estimation is done to shed light on accuracy and possible flaws and uncertainties of the risk analysis. The direct economic loss from the 2011 Tohoku tsunami was reported to be 210 Billion USD, while the 2004 Indian Ocean tsunami caused an economic loss of 4.4 Billion USD (www.emdat.be). For Japan, the PML curve gives a loss of 340 Billion USD for a mean return period of 500 years while a loss equal to 210 Billion USD (the same as given in www.emdat.be) is expected for a mean return period of 250 years. For Indonesia, an economic loss of 4.5 Billion USD is obtained for a mean

return period of 500 years. Return periods for giant megathrust earthquakes for Japan are uncertain, but ballpark estimates have indicated 500 years and upward (Kagan and Jackson 2013), which indicate that the calculated PML agrees closely with the observations for Japan given the various uncertainties at play. For Indonesia, we are not able to make a similar direct comparison but observed that the computed PML for a mean return period of 500 years has the same order of magnitude and is comparable to the loss recorded after the 2004 Indian Ocean tsunami. We know that paleotsunami evidence (Jankaew et al. 2008; Monecke et al. 2008) loosely supports a cycle of about a 1,000 years or more (order of magnitude) for events similar to the 2004 Indian Ocean Tsunami based on sediment deposits. However, we do not know how the magnitude of the sources that gave rise to these paleotsunamis compares with the M_w 9.1 2004 Great Sumatra Earthquake, nor the maximum run-up induced by these tsunamis.

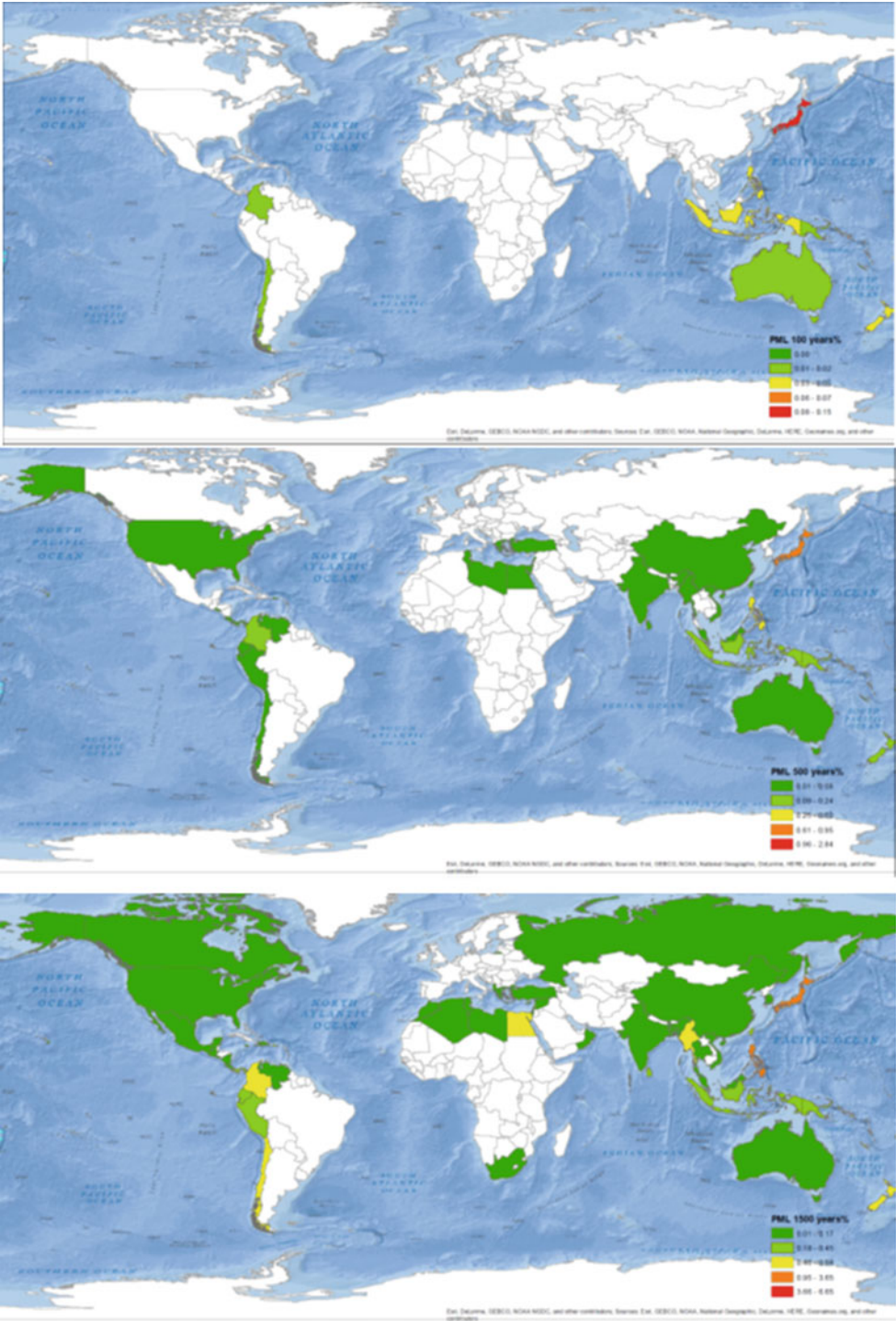
A Note About Simplifying Assumptions and Uncertainties

The global nature of this study involves relatively large uncertainties. Some uncertainties (aleatory) are included through the standard deviation σ_H related to the tsunami flow depth, thereby being considered and rigorously propagated into the tsunami hazard and risk assessment. Other uncertainties may involve systematic deviations that are less easily captured through the aleatory uncertainty formulation. As a consequence of the rough nature of the different methods involved, some bias is therefore expected in addition to the modeled uncertainty. The most conservative assumptions (toward overestimating tsunami probabilities) are linked to the tectonic



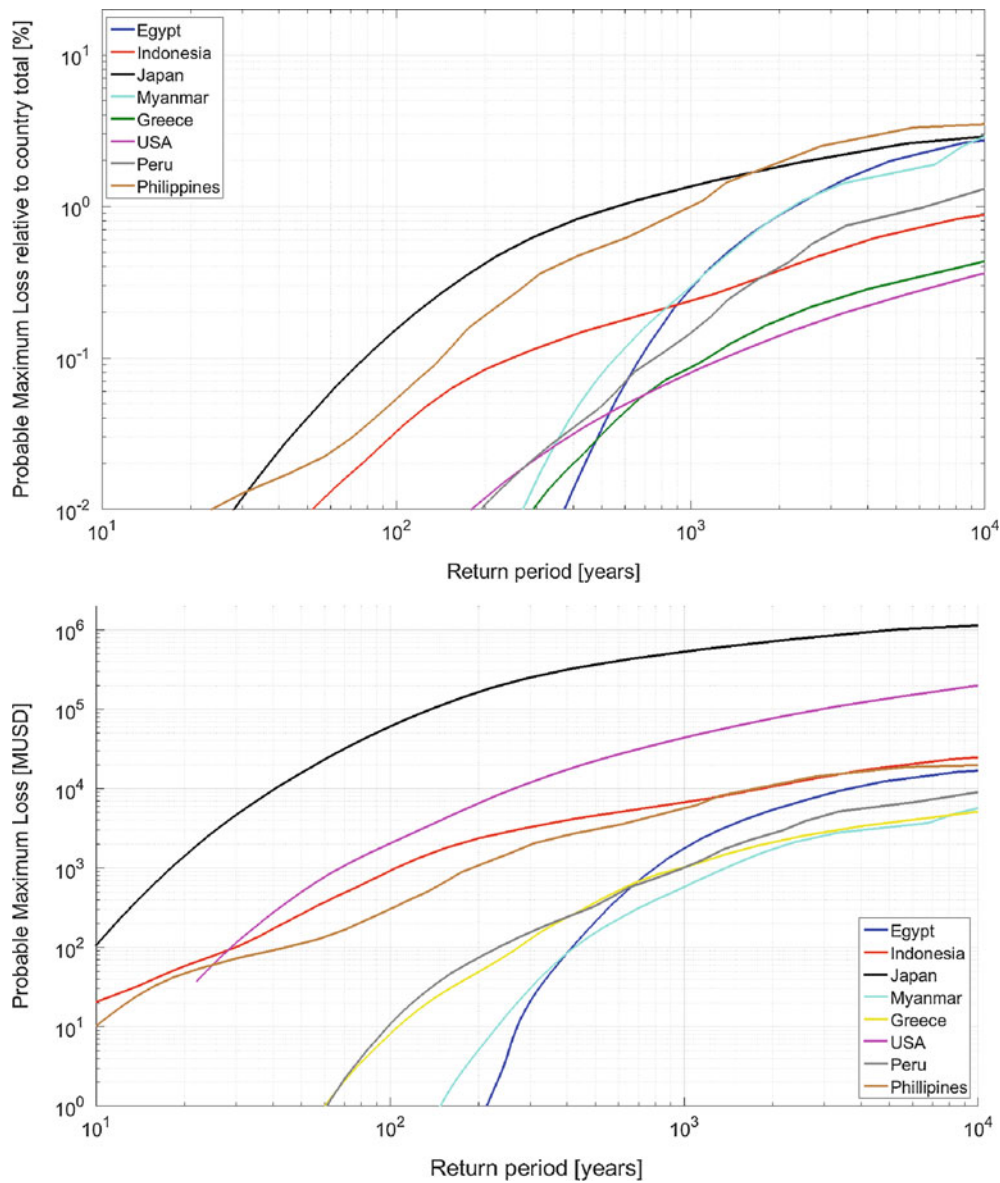
Tsunami Hazard and Risk Assessment on the Global Scale, Fig. 13 Probable Maximum Losses (in USD Million) for all countries worldwide for return periods of 100, 500, and 1,500 years. These losses are only related to direct economic losses from damage to buildings. *White*

areas denotes low, or no values of loss. Note that the color scales are different between the different panels (CIMNE and INGENIAR 2015). Note that very limited number of countries, such as Madagascar, were missing in this analysis.



Tsunami Hazard and Risk Assessment on the Global Scale, Fig. 14 Probable Maximum Losses, normalized by the total direct economic value of buildings, for all countries worldwide for return periods of 100, 500, and 1,500 years. These losses are only related to direct

economic losses from damage to buildings. *White* areas denotes low, or no values of loss. Note that the color scales are different between the different panels (CIMNE and INGENIAR 2015). Note that very limited number of countries, such as Madagascar, were missing in this analysis.



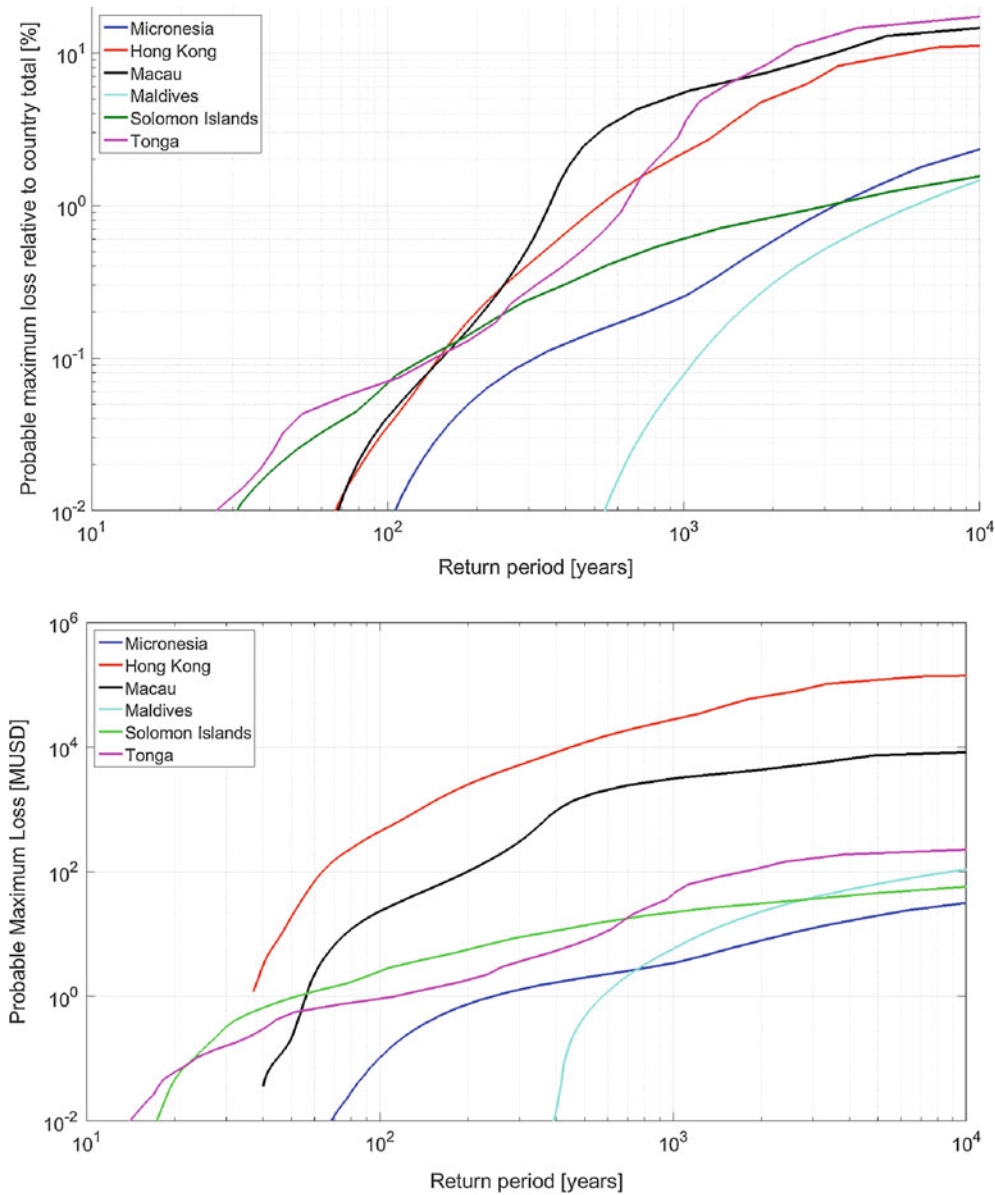
Tsunami Hazard and Risk Assessment on the Global Scale, Fig. 15 Selected PML curves for a selection of moderately large to large countries. *Upper panel*, loss relative to the total exposed value for the whole country.

Lower panel, probable maximum loss in USD Million. Missing data points indicate negligible losses (associated with short return periods)

assumptions, whereas the most underconservative assumptions (toward underestimating tsunami exposure and risk) stem from the treatment of the inundation. We expect that the latter is the larger of the two. The various uncertainties and limitations of the study are elaborated further below.

Source Representation

The PTHA method incorporates a weighting of different tectonic and seismicity information used to determine the scenario parameters and their associated mean return periods. The weighting involves expert judgment to account for the need to model events where supporting



Tsunami Hazard and Risk Assessment on the Global Scale, Fig. 16 Selected PML curves for a selection of small island countries. *Upper panel*, loss relative to the total exposed value for the whole country. *Lower panel*,

probable maximum loss in USD Million. Missing data points indicate negligible losses (associated with short return periods)

data for the recurrence is either scarce or nonexistent. As a consequence, full seismic coupling is assumed in the present analysis, which is certainly conservative. We expect the uncertainty related to the occurrence of large events to be particularly prominent, and in general the resulting return periods are clearly subject to

large uncertainty. The uncertainty σ_R is set to a single value but will in general vary from region to region, which is another simplification. Moreover, the applied lognormal distribution implies finite probabilities for a large flow depth at any inundated point. While the lognormal distribution was chosen for compatibility with the risk

estimation, it is clear that the tsunami flow depth is expected to be bounded. Another source of uncertainty is related to the shear modulus μ that enters the seismic moment and therefore indirectly affects the mean return period. In particular, the shear modulus at shallow depth has large influence on the tsunamigenesis, and low values of μ imply more efficient wave generation. The shear modulus is clearly subject to variability (see, e.g., Bilek and Lay 1999). A proper statistical treatment of the so-called tsunami earthquakes (see, e.g., Kanamori 1972, for a discussion) associated with low values of μ was not included herein. Many important examples that are probable or almost certain for slow events include Sanriku 1896, Aleutians 1946, Nicaragua 1992, Java 1994, Java 2006, and Mentawai 2010. Most of these events have magnitudes in the lower end of those considered in our analysis.

It should also be noted that tsunamis generated by nonseismic sources such as volcanoes and submarine and subaerial landslides are not considered in the present study. Nonseismic sources contribute to the generation of about one fifth of all tsunamis globally (Harbitz et al. 2014a), and there are several examples of such tsunamis causing devastation. A recent example is the 1998 Papua New Guinea tsunami caused by a submarine landslide triggered by an earthquake, causing 2,182 fatalities (<http://www.emdat.be>). In areas like eastern Indonesia (Løvholt et al. 2012b), the Mediterranean (Tinti et al. 2005), and the Caribbean (Harbitz et al. 2012) tsunamis due to landslides and volcanoes are relatively more frequent and contribute to a significant portion of the risk. Therefore, the tsunami hazard may be underestimated in such areas where landslides are more frequent if compared with events triggered by earthquakes. In Norway, where virtually all tsunamis are caused by landslides (Harbitz et al. 2009, 2014b), the derived hazard maps based on earthquakes as triggering events are not representative. Unlike earthquakes, landslides are not constrained to the major subduction zones and may strike unexpectedly. Due to their source characteristics, landslides and volcanoes may generate larger run-up locally compared to earthquakes but propagate less efficiently to the

far-field (Harbitz et al. 2014a). However, addressing their return periods involves a larger scale of uncertainty than for earthquakes.

Exposure and Risk

The methods for establishing the hazard maps and population exposure are approximate and simplified in order to cover large geographic areas given the global scope of the assessment. Inundation maps are established by intersecting extrapolated run-ups on coarse topographic data (SRTM), which is a major simplification. SRTM data (Rodriguez et al. 2005) has a vertical accuracy that may often include areas of falsely elevated land, and may reduce the inundated area to yield artificially low inundation areas and flow depths. Griffin et al. (2015) showed how this may lead to a large underestimation of the inundation footprint and consequently an underestimation of the exposed elements. Løvholt et al. (2014a) demonstrated that the sensitivity due to moderate changes (+2 m elevation change) in the SRTM may be large (see Table 3). Knowing that the SRTM data have a particularly positive bias in tropical and urban areas (Kaiser et al. 2011; Römer et al. 2012), we conclude that the present exposure based on SRTM has lower-bound estimates. Moreover, the effects of countermeasures such as breakwaters, which are expected to decrease the risk by lowering the exposure, are not considered herein. Breakwaters are, for instance, common in Japan. In the hazard maps, differences in the reference location of the coastline sections sometimes occur. These differences

Tsunami Hazard and Risk Assessment on the Global Scale, Table 3 Population exposure from GAR13 (500 years return period) for four different countries. The population exposure when uniformly subtracting 2 m from the topographic SRTM data is compared with the population exposure using pure SRTM data. Results are extracted from Løvholt et al. (2014a)

	SRTM	SRTM – 2 m
Myanmar	253,000	312,000
Sri-Lanka	209,000	335,000
Pakistan	24,000	49,000
India	529,000	851,000

may cause slight offsets between the affected zones and coastlines.

Regarding the specific risk metrics (AAL and PML) and their associated uncertainties, it is important to mention that, for example, since AAL represents the loss results in annual terms, values which correspond mathematically to an expected value cannot have any associated uncertainty measure. For the PML, the uncertainty is considered in the calculation of the exceedance probabilities (a probability calculated for a loss value) and, therefore, the obtained annual exceedance rates cannot have any associated uncertainties (Bernal 2014; Salgado-Gálvez et al. 2015).

Considering the uncertainties as well as the correlation of losses in a proper manner, such as in the methodology proposed by Ordaz (2000), has effects in the initial part of the PML plot (Bazzurro and Luco 2005) with the most frequent events having a strong influence on the final AAL value. Altogether, it has been determined that uncertainties have less impact on the AAL than on the complete LEC (Marulanda 2013).

Future Directions

In order to improve our present understanding of tsunami hazard and risk, aspects related to the uncertainties and inaccuracies listed above must be addressed. As stated, major uncertainties exist in our treatment of the sources, inundation and exposure, and our understanding of the vulnerability of people and assets. All of these factors need to be given considerable attention in future research. Certain factors, such as quantifying the inundation and exposure, have been limited due to feasibility reasons and may be improved by using more accurate methods and by including more accurate topography data. Our understanding of probabilistic tsunami hazard and exposure may be improved through a bottom-up approach that integrates the work of a wider tsunami community or by calibrating the amplification method to a greater extent. However, key factors such as tsunami recurrence continue to be subject to large epistemic uncertainty due to our lack of

data. While some of this uncertainty may be reduced in the future, improving hazard assessment at present will depend on more complete formal consideration of epistemic and aleatory uncertainty in tsunami source models. To this end, there is presently much ongoing research on improving our understanding of uncertainties related to earthquake fault parameters and slip variability. Synergies between efforts on seismic hazard studies should also be exploited. We also expect our understanding of tsunami vulnerability to be improved as more data is collected from future tsunamis. While tsunami risk assessment is considered a fairly young discipline, with a limited literature at present, the steadily increasing number of papers in this field may suggest that we will see more sophisticated approaches in the future.

Acknowledgments A large group of scientists have worked on the global tsunami hazard and risk assessment. For this, we would like to thank our co-workers at Geoscience Australia, NGI, and CIMNE: Gareth Davies, Sylfest Glimsdal, Carl Harbitz, Helge Smebye, Farrokh Nadim, Omar Darío Cardona, and Gabriel A. Bernal. In addition we have received considerable support from collaborating organizations. For this we thank Andrea de Bono at UNEP/GRID Geneva for providing us the exposure dataset and for refining them for use in coastal regions, Stefano Lorito, Roberto Basili, and Jacopo Selva at INGV for providing source information for the Mediterranean, Maria Ana Baptista at IPMA for providing source information offshore Portugal, Eric Geist at USGS for providing source information for the Caribbean region, and Hong Kie Thio at AECOM for his assistance on the PTHA calculations. The authors are also indebted to Nick Horspool, formerly at Geoscience Australia, now at GNS, for his contribution to the earliest part of this work. We finally like to thank UN-ISDR, including Andrew Maskrey, Sahar Safaie, Manuela di Mauro, and Julio Serje, for their co-ordination work that made this possible, as well as for funding NGI's work leading to this paper. This paper is published with the permission of the CEO, Geoscience Australia.

Bibliography

- Annaka T, Satake K, Sakakiyama T, Yanagisawa K, Shuto N (2007) Logic-tree approach for probabilistic tsunami hazard analysis and its applications to the Japanese coasts. *Pure Appl Geophys* 164:577–592
- Barkan R, ten Brink U, Lin J (2009) Far field tsunami simulations of the 1755 Lisbon earthquake:

- implications for tsunami hazard to the U.S. East Coast and the Caribbean. *Mar Geol* 264(1–2):109–122
- Basili R, Tiberti MM, Kastelic V, Romano F, Piatanesi A, Selva J, Lorito S (2013) Integrating geologic fault data into tsunami hazard studies. *Nat Hazards Earth Syst Sci* 13:1025–1050
- Bazzurro P, Luco N (2005) Accounting for uncertainty and correlation in earthquake loss estimation. In: Proceedings of ICOSSAR, Rome, pp. 2687–2694
- Bernal GA (2014) Metodología para la modelación, cálculo y calibración de parámetros de la amenaza sísmica para la evaluación probabilista del riesgo (in Spanish). PhD thesis, Technical University of Catalonia, Barcelona
- Berryman K et al (ed) (2005) Review of tsunami hazard and risk in New Zealand. Geological and Nuclear Sciences (GNS), report 2005/104, 140 p
- Berryman K, Wallace L, Hayes G, Bird P, Wang K, Basili R, Lay T, Stein R, Sagiya T, Rubin C, Barreiros S, Kreemer C, Litchfield N, Pagani M, Gledhill K, Haller K, Costa C (2013) The GEM faulted earth subduction characterisation project, version 1.0, June 2013. <http://www.nexus.globalquakemodel.org/gem-faulted-earth/posts>
- Bilek S and Lay T (1999) Rigidity with depth along interpolate megathrust faults in subduction zones, *Nature* 400, 443–446. <https://doi.org/10.1038/22739>
- Bird P (2003) An updated digital model of plate boundaries. *Geochem Geophys Geosyst* 4(3):1027. <https://doi.org/10.1029/2001GC000252>
- Borrero J, Sieh K, Chlieh M, Synolakis C (2006) Tsunami inundation modelling for western Sumatra. *Proc Natl Acad Sci U S A* 103(52):19673–19677
- Brizuela B, Armigliato A, Tinti S (2014) Assessment of tsunami hazards for the Central American Pacific coast from southern Mexico to northern Peru. *Nat Hazards Earth Syst Sci* 14:1889–1903. <https://doi.org/10.5194/nhess-14-1889-2014>
- Burbridge D, Cummins P (2007) Assessing the threat to Western Australia from tsunami generated by earthquakes along the Sunda Arc. *Nat Hazards* 43:319–331. <https://doi.org/10.1007/s11069-007-9116-3>
- Burbridge D, Cummins PR, Mleczko R, Thio HK (2008a) A probabilistic tsunami hazard assessment for Western Australia. *Pure Appl Geophys.* <https://doi.org/10.1007/s00024-008-0421-x>
- Burbridge D, Mleczko R, Thomas C, Cummins P, Nielsen O, Dhu T (2008b) A probabilistic tsunami hazard assessment for Australia. *Geoscience Australia Professional Opinion*. No.2008/04
- Burroughs SF, Tebbens SM (2005) Power-law scaling and probabilistic forecasting of tsunami runup heights. *Pure Appl Geophys* 162:331–342
- Cardona OD (2009) La gestión financiera del riesgo de desastre. Instrumentos financieros de retención y transferencia para la Comunidad Andina (in Spanish). PREDECAN, Lima
- Cardona OD, Ordaz M, Reinoso E, Yamín LE, Barbat AH (2012) CAPRA – comprehensive approach to probabilistic risk assessment: international initiative for risk management effectiveness. In: Proceedings of 15th world conference on earthquake engineering, Lisbon
- Cardona OD, Ordaz M, Mora MG, Salgado-Gálvez MA, Bernal GA, Zuloaga D, Marulanda MC, Yamín LE, González D (2014) Global risk assessment: a fully probabilistic seismic and tropical cyclone wind risk assessment. *Int J Disaster Risk Reduct* 10:461–476
- Carrier GF, Greenspan HP (1958) Water waves of finite amplitude on a sloping beach. *J Fluid Mech* 4:97–109
- CIMNE, INGENIAR (2015) Update on the probabilistic global of natural risk at the global level: global risk model. Background paper for the global assessment report on disaster risk reduction 2015. http://www.preventionweb.net/english/hyogo/gar/2015/en/home/documents.html#contributing_papers
- CIMNE, ITEC, INGENIAR, EAI (2013) Probabilistic modelling of natural risks at the global level. Global risk model. Background paper for the global assessment report on disaster risk reduction 2013. http://www.preventionweb.net/english/hyogo/gar/2013/en/home/documents.html#contributing_papers
- Cornell CA (1968) Engineering seismic risk analysis. *Bull Seismol Soc Am* 58(5):1583–1606
- De Bono A, Chatenoux B (2014) A global exposure model for GAR 2015. Input paper for the global assessment report on disaster risk reduction 2015. http://www.preventionweb.net/english/hyogo/gar/2015/en/home/documents.html#contributing_papers
- Dominey-Howes D, Papathoma M (2007) Validating a tsunami vulnerability assessment model (the PTVA Model) using field data from the 2004 Indian Ocean tsunami. *Nat Hazards* 40:113–36
- Dominey-Howes D, Dunbar P, Varner J, Papathoma-Köhle M (2010) Estimating probable maximum loss from a Cascadia tsunami. *Nat Hazards* 53(1):43–61
- Esteva L (1967) Criterios para la construcción de espectros de diseño sísmico (in Spanish). In: Proceedings of the 3rd pan-American symposium of structures, Caracas
- Geist E, Parsons T (2006) Probabilistic analysis of tsunami hazards. *Nat Hazards* 37:277–314
- Gonzalez FI, Geist EL, Jaffe B et al (2009) Probabilistic tsunami hazard assessment at Seaside, Oregon, for near- and far-field seismic sources. *J Geophys Res Oceans* 114:C11023
- Griffin JG, Latief H, Kongko W, Harig S, Horspool N, Hanung R, Rojali A, Maher N, Fountain L, Fuchs A, Hossen J, Upi S, Dewanto SE, Cummins PR (2015) An evaluation of onshore digital elevation models for modelling tsunami inundation zones. *Front Earth Sci* 3:32. <https://doi.org/10.3389/feart.2015.00032>
- Grilli ST, Dubosq S, Popinet N, Pérignon Y, Kirby JT, Shi F (2010) Numerical simulation and first-order hazard analysis of large co-seismic tsunamis generated in the Puerto Rico trench: near-field impact on the North shore of Puerto Rico and far-field impact on the US East Coast. *Nat Hazards Earth Syst Sci* 10:2109–2125. <https://doi.org/10.5194/nhess-10-2109-2010>
- Harbitz CB, Glimsdal S, Løvholt F, Pedersen GK, Vanneste M, Eidsvig UMK, Bungum H (2009)

- Tsunami hazard assessment and early warning systems for the North East Atlantic. In: Proceedings of DEWS midterm conference, Potsdam, 7–8 July 2009. http://www.dews-conference.org/front_content.php
- Harbitz CB, Glimsdal S, Bazin S, Zamora N, Løvholt F, Bungum H, Smebye H, Gauer P, Kjekstad O (2012) Tsunami hazard in the Caribbean: regional exposure derived from credible worst case scenarios. *Cont Shelf Res* 38:1–23
- Harbitz CB, Løvholt F, Bungum H (2014a) Submarine landslide tsunamis: how extreme and how likely? *Nat Hazards* 72(3):1341–1374
- Harbitz CB, Glimsdal S, Løvholt F, Kvelodsvik V, Pedersen GK, Jensen A (2014b) Rockslide tsunamis in complex fjords: from an unstable rock slope at Åkerneset to tsunami risk in western Norway. *Coast Eng* 88:101–122. <https://doi.org/10.1016/j.coastaleng.2014.02.003>
- Hayes GP, Wald DJ, Johnson RL (2012) Slab1.0: a three-dimensional model of global subduction zone geometries. *J Geophys Res Solid Earth* (1978–2012) 117(B1):2156–2202
- Hebert H, Schindele F, Altinok Y, Alpar B, Gazioglu C (2005) Tsunami hazard in the Marmara Sea (Turkey): a numerical approach to discuss active faulting and impact on the Istanbul coastal areas. *Mar Geol* 215:23–43
- Heidarzadeh M, Kijko A (2011) A probabilistic tsunami hazard assessment for the Makran subduction zone at the Northwestern Indian Ocean. *Nat Hazards* 56(3):577–593
- Hooper A, Pietrzak J, Simons W, Cui H, Riva R, Naeije M, Terwisscha van Scheltinga A, Schrama E, Stelling G, Socquet A (2013) Importance of horizontal seafloor motion on tsunami height for the 2011 Mw = 9.0 Tohoku-Oki earthquake. *Earth Planet Sci Lett* 361(1):469–479
- Horspool N, Pranantyo I, Griffin J, Latief H, Natawidjaja DH, Kongko W, Cipta A, Bustaman B, Anugrah SD, Thio HK (2014) A probabilistic tsunami hazard assessment for Indonesia. *Nat Hazards Earth Syst Sci* 14(11):3105–3122
- Jankaew K, Atwater BF, Sawai Y, Choowong M, Charoentitrat T, Martin ME, Prendergast A (2008) Medieval forewarning of the 2004 Indian Ocean tsunami in Thailand. *Nature* 455:1228–1231
- Kagan YY, Jackson DD (2013) Tohoku earthquake: a surprise? *Bull Seismol Soc Am* 103(2B):1181–94
- Kaiser G, Scheele L, Kortenhaus A, Løvholt F, Römer H, Leschka S (2011) The influence of land cover roughness on high resolution tsunami inundation modeling. *Nat Hazards Earth Syst Sci* 11:2521–2540
- Kaistrenko VM, Pinegina T, Klyachko MA (2003) Evaluation of tsunami hazard for the southern Kamchatka coast using historical and paleotsunami data. In: Yalciner AC, Pelinovsky E, Okal E, Synolakis CE (eds) *Submarine landslides and tsunamis*. Kluwer, Dordrecht, pp 217–228
- Kanamori H (1972) Mechanisms of tsunami earthquakes. *Phys Earth Planet Inter* 6:346–359
- Kramer SL (1996) *Geotechnical earthquake engineering*, Prentice-Hall international series in civil engineering and engineering mechanics. Prentice Hall, Upper Saddle River
- Kulikov EA, Rabinovich AB, Thomson RE (2005) Estimation of tsunami risk for the coasts of Peru and northern Chile. *Nat Hazards* 35:185–209
- LandScan (2013) High resolution global population data set © UT-Battelle, LLC, operator of Oak Ridge National Laboratory, USA. Dataset is available upon demand to ONRL
- Lane EM, Gillibrand PA, Wang XA (2013) Probabilistic tsunami hazard study of the Auckland region, Part II: inundation modelling and hazard assessment. *Pure Appl Geophys* 170(9–10):1635–1646
- Laske G, Masters G, Ma Z, Pasyanos ME (2012) CRUST1.0: an updated global model of earth's crust. In: Proceedings of Europ Geosciences Union General Assembly 2012, Vienna
- Legg MR, Borrero JC, Synolakis CE (2004) Tsunami Hazards associated with the Catalina fault in Southern California. *Earthquake Spectra* 20(3):917–950
- Lin I-C, Tung CC (1982) A preliminary investigation of tsunami hazard. *Bull Seismol Soc Am* 72(A):2323–2337
- Liu Y, Santos A, Wang SM, Shi Y, Liu H, Yuen DA (2007) Tsunami hazards along Chinese coast from potential earthquakes in South China Sea. *Phys Earth Planet Inter* 163:233–244
- Lorito S, Tiberti MM, Basili R, Piatanesi A, Valensise G (2008) Earthquake-generated tsunamis in the Mediterranean Sea: scenarios of potential threats to Southern Italy. *J Geophys Res Solid Earth* 113(B1):B01301
- Lorito S, Selva J, Basili R, Romano F, Tiberti MM, Piatanesi A (2015) Probabilistic hazard for seismically induced tsunamis: accuracy and feasibility of inundation maps. *Geophys J Int* 200:574–588
- Løvholt F, Bungum H, Harbitz CB, Glimsdal S, Lindholm CD, Pedersen G (2006) Earthquake related tsunami hazard along the western coast of Thailand. *Nat Hazards Earth Syst Sci* 6:1–19
- Løvholt F, Pedersen G, Gisler G (2008) Oceanic propagation of a potential tsunami from the La Palma Island. *J Geophys Res Oceans* 113:C09026. <https://doi.org/10.1029/2007JC004603>
- Løvholt F, Glimsdal S, Harbitz CB, Nadim F, Zamora N, Peduzzi P, Dao HI, Smebye H (2012a) Tsunami hazard and exposure on the global scale. *Earth-Sci Rev* 110(1–4):58–73. <https://doi.org/10.1016/j.earscirev.2011.10.002>, ISSN 0012–8252
- Løvholt F, Kühn D, Bungum H, Harbitz CB, Glimsdal S (2012b) Historical tsunamis and present tsunami hazard in Eastern Indonesia and the Philippines. *J Geophys Res Solid Earth* 117:B09310. <https://doi.org/10.1029/2012JB009425>

- Løvholt F, Lynett P, Pedersen G (2013) Simulating run-up on steep slopes with operational Boussinesq models; capabilities, spurious effects and instabilities. *Nonlinear Process Geophys* 20:379–395. <https://doi.org/10.5194/npg-20-379-2013>
- Løvholt F, Glimsdal S, Harbitz CB, Horspool N, Smebye H, de Bono A, Nadim F (2014a) Global tsunami hazard and exposure due to large co-seismic slip. *Int J Disaster Risk Reduct* 10:406–418
- Løvholt F, Setiadi NJ, Birkmann J, Harbitz CB, Bach C, Fernando N, Kaiser G, Nadim F (2014b) Tsunami risk reduction – are we better prepared today than in 2004? *Int J Disaster Risk Reduct* 10:127–142
- Maqsood T, Wehner M, Ryu H, Edwards M, Dale K, Miller V (2014) GAR15 regional vulnerability functions. *Geoscience Australia Record* 2014/38
- Marulanda MC (2013) Modelación probabilista de pérdidas económicas por sismo para la estimación de la vulnerabilidad fiscal del estado y la gestión financiera del riesgo soberano. PhD thesis (in Spanish), Technical University of Catalonia, Barcelona
- Matias LM, Cunha T, Annunziato A, Baptista MA, Carrilho F (2013) Tsunamigenic earthquakes in the Gulf of Cadiz: fault model and recurrence. *Nat Hazards Earth Syst Sci* 13:1–13. <https://doi.org/10.5194/nhess-13-1-2013>
- Mercado A (2001) Determination of the tsunami hazard for western Puerto Rico from local sources. Sea Grant College Program University of Puerto Rico, P.R. (Report)
- Miranda E (1999) Approximate seismic lateral deformation demands in multistory buildings. *J Struct Eng* 125:417–426
- Monecke K, Finger W, Klarer D, Kongko W, McAdoo BG, Moore AL, Sudrajat SU (2008) A 1,000-year sediment record of tsunami recurrence in northern Sumatra. *Nature* 455:1232–1234
- NGDC/WDS Global Historical Tsunami Database: https://www.ngdc.noaa.gov/hazard/tsu_db.shtml
- NGI and Geoscience Australia (2015) UNISDR global assessment report 2015 – GAR15, Tsunami methodology and result overview. NGI report 20120052-03-R
- Okada Y (1985) Surface deformation due to shear and tensile faults in a half-space. *Bull Seismol Soc Am* 74(4):1135–1154
- Okal E, Synolakis CE (2008) Far-field tsunami hazard from mega-thrust earthquakes in the Indian Ocean. *Geophys J Int* 172:995–1015
- Okal EA, Borrero JC, Synolakis CE (2006) Evaluation of tsunami risk from regional earthquakes at Pisco, Peru. *Bull Seismol Soc Am* 96(5):1634–1648
- Okal EA, Synolakis CE, Kalligeris N (2011) Tsunami simulations for regional sources in the South China and adjoining Seas. *Pure Appl Geophys* 168(6–7):1153–1173
- Omira R, Baptista MA, Matias L (2014) Probabilistic tsunami hazard in the Northeast Atlantic from near- and far-field tectonic sources. *Pure Appl Geophys* 172:901–920
- Ordaz M (2000) Metodología para la evaluación del riesgo sísmico enfocada a la gerencia de seguros por terremoto (in Spanish). Universidad Nacional Autónoma de México, Mexico City
- Ordaz M, Miranda E, Reinoso E, Pérez-Rocha LE (2000) Seismic loss estimation model for Mexico City. In: *Proceedings of the 12th world conference on earthquake engineering*, Auckland
- Ozawa S, Nishimura T, Suito H, Kobayashi T, Tobita M, Imakiire T (2011) Coseismic and postseismic slip of the 2011 magnitude-9 Tohoku-Oki earthquake. *Nature* 475:373–376. <https://doi.org/10.1038/nature10227>
- Parsons T, Geist E (2009) Tsunami probability in the Caribbean region. *Pure Appl Geophys* 165: 2089–2116
- Pedersen G (2008) Modeling run-up with depth integrated equation models. In: Liu PL-F, Yeh H, Synolakis C (eds) *Advanced numerical models for simulating tsunami waves and run-up*. World Scientific, Hackensack, pp 3–41
- Pedersen G (2011) Oblique runup of non-breaking solitary waves on an inclined plane. *J Fluid Mech* 668:582–606. <https://doi.org/10.1017/S0022112010005343>
- Pedersen G, Løvholt F (2008) Documentation of a global Boussinesq solver, Preprint series in applied mathematics, Department of Mathematics, University of Oslo, Norway. <http://urn.nb.no/URN:NBN:no-27775>
- Rodriguez E, Morris CS, Belz JE, Chapin EC, Martin JM, Daffer W, Hensley S (2005) An assessment of the SRTM topographic products. Jet-Propulsion Laboratory D-31639. http://www2.jpl.nasa.gov/srtm/SRTM_D31639.pdf
- Roger J, Hebert H (2008) The 1856 Djijelli (Algeria) earthquake and tsunami: source parameters and implications for tsunami hazard in the Balearic Islands. *Nat Hazards Earth Syst Sci* 8:721–731
- Römer H, Willroth P, Kaiser G, Vafeidis AT, Ludwig R, Sterr H, Revilla Diez J (2012) Potential of remote sensing techniques for tsunami hazard and vulnerability analysis – a case study from Phang-Nga province, Thailand. *Nat Hazards Earth Syst Sci* 12:2103–2126
- Salgado-Gálvez MA, Zuloaga D, Bernal GA, Mora MG, Cardona OD (2014) Fully probabilistic seismic risk assessment considering local site effects for the portfolio of buildings in Medellín, Colombia. *Bull Earthq Eng* 12:671–695
- Salgado-Gálvez MA, Zuloaga D, Velásquez CA, Carreño ML, Cardona OD, Barbat AH (2015a) Urban seismic risk index for Medellín, Colombia, based on probabilistic loss and casualties estimations. *Nat Hazards*. DOI: 10.1007/s11069-015-2056-4. (In press)
- Salgado-Gálvez MA, Cardona OD, Carreño ML, Barbat AH (2015b) Probabilistic seismic hazard and risk assessment in Spain. *Monographs on earthquake engineering*. CIMNE, Barcelona. <https://doi.org/10.13140/2.1.3073.1049>

- Satake K (1995) Linear and non-linear computations of the 1992 Nicaragua earthquake tsunami. *Pure Appl Geophys* 144:455–470
- Sørensen MB, Spada M, Babeyko A, Wiemer S, Grünthal G (2012) Probabilistic tsunami hazard in the Mediterranean Sea. *J Geophys Res* 117(B1):2156–2202. <https://doi.org/10.1029/2010JB008169>
- Stein S, Okal EA (2005) Speed and size of the Sumatra earthquake. *Nature* 434:581–582
- Stein S, Okal EA (2007) Ultralong period seismic study of the December 2004 Indian Ocean earthquake and implications for regional tectonics and the subduction process. *Bull Seismol Soc Am* 97(1A):S279–S295. <https://doi.org/10.1785/0120050617>
- Strasser FO, Arango MC, Bommer JJ (2010) Scaling of the source dimensions of interface and intraslab subduction-zone earthquakes with moment magnitude. *Seismol Res Lett* 81(6):941–950
- Suppasri A, Imamura F, Koshimura S (2012) Probabilistic tsunami hazard analysis and risk to coastal populations in Thailand. *J Earthq Tsunami* 6(2). <https://doi.org/10.1142/S179343111250011X>
- Suppasri A, Mas E, Charvet I, Gunasekera R, Imai K, Fukutani Y, Abe Y, Imamura F (2013) Building damage characteristics based on surveyed data and fragility curves of the 2011 Great East Japan tsunami. *Nat Hazards* 66(2):319–341. <https://doi.org/10.1007/s11069-012-0487-8>
- Synolakis CE, Bernard EN, Titov VV, Kânoglu U, González F (2007) Validation and verification of tsunami numerical models. *Pure Appl Geophys* 165:2197–2228
- Tadepalli S, Synolakis CE (1996) Model for the leading waves of tsunamis. *Phys Rev Lett* 77(10):2141–2144
- ten Brink U (2005) Vertical motions of the Puerto Rico Trench and Puerto Rico and their cause. *J Geophys Res* 110:B06404. <https://doi.org/10.1029/2004JB003459>
- Thio HK, Somerville P, Polet J (2010) Probabilistic tsunami hazard in California, PEER report 2010/108 Pacific Earthquake Engineering Research Center
- Tinti S, Armigliato A (2003) The use of scenarios to evaluate the tsunami impact in southern Italy. *Mar Geol* 199(3–4):221–243
- Tinti S, Manucci A, Pagnoni G, Armigliato A, Zaniboni F (2005) The 30 December 2002 landslide-induced tsunamis in Stromboli: sequence of the events reconstructed from the eyewitness accounts. *Nat Hazards Earth Syst Sci* 5(6):763–775
- Titov VV, Gonzalez FI (1997) Implementation and testing of the Method of Splitting Tsunami (MOST) model. NOAA. Technical Memorandum ERL PMEL-112, 11 pp
- Titov VV, Rabinovich AB, Mofjeld HO, Thomson RE, Gonzalez FI (2005) The global reach of the 26 December 2004 Sumatra tsunami. *Science* 309(5743):2045–2048
- Tsunami Laboratory Novosibirsk, Historical Tsunami Database for the World Ocean (HTDB/WLD), <http://tsun.sscc.ru/>
- UN-ISDR (2015) GAR – global assessment report on disaster risk reduction – making development sustainable: the future of disaster risk management. Report. Available from www.preventionweb.net/gar/
- Venturato AJ, Arcas D, Kanoglu U (2007) Modeling tsunami inundation from a Cascadia subduction zone earthquake for long beach and Ocean Shores, Washington. NOAA technical memorandum OAR PMEL-137. U.S. Department of Commerce, National Oceanic and Atmospheric Administration, Office of Oceanic and Atmospheric Research, Pacific Marine Environmental Laboratory, Seattle, pp 13
- Wang R, Martín FL, Roth F (2003) Computation of deformation induced by earthquakes in a multi-layered elastic crust – FORTRAN programs EDGRN/EDCMP. *Comput Geosci* 29(2):195–207

Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data

Stefano Lorito¹, F. Romano² and T. Lay³

¹Istituto Nazionale di Geofisica e Vulcanologia, Rome, Italy

²Roma 1, Sez. Sismologia e Tettonofisica, Istituto Nazionale di Geofisica e Vulcanologia, Roma, Italy

³Department of Earth and Planetary Sciences, University of California Santa Cruz, Santa Cruz, CA, USA

Article Outline

Glossary
Definition of the Subject
Introduction
Tsunamis and the Seismic Source
Inversion for the Seismic Source
Megathrust Events
Great Subduction Earthquake Doublets
Strike-Slip Events
Tsunami Earthquakes
Special Cases
Future Directions
Bibliography

Glossary

Earthquake Location and Fault Parameters

Dip The angle between the Earth's horizontal surface and the **earthquake fault**

Epicenter The position (longitude, latitude) of the **hypocenter** projected on the Earth's surface

Earthquake Fault The fracture within a rock volume that separates two rock masses where there is evidence of a relative displacement (**slip**) between them during an earthquake; the surface where the slip occurred is termed the **fault surface**; if the fault surface is inclined, the rock masses above and below the fault are termed **hanging wall** and **footwall**, respectively

Hypocenter The position (longitude, latitude, depth) of the earthquake nucleation point

Rake (or Slip Direction) The angle of the relative displacement between the hanging wall and the footwall of a fault during an earthquake measured in the fault surface relative to the horizontal **strike** direction

Strike The angle, measured relative to north, of the intersection (**trace**) of the earthquake fault with the Earth's surface

Faulting Style

Normal Fault Fault where the hanging wall moves down with respect to the footwall

Reverse Fault Fault where the hanging wall moves up with respect to the footwall; if the dip angle is $\leq 45^\circ$, it is termed a **thrust fault**

Strike-Slip Fault Fault (often steeply dipping) where the two blocks separated by the fault surface move horizontally along the fault trace relative to each other; if the motion of the block across the fault trace from an observer moves toward left, the mechanism is termed **left lateral**, otherwise the mechanism is termed **right lateral**

Transpressive Fault Reverse fault with an oblique rake component

Kinematic Rupture Parameters and Inversion

Inversion (or Inverse Problem) A procedure that uses a set of direct or indirect observations to retrieve the parameters of a model that is consistent with the observations; this is a common approach (many techniques are used

depending on the problem to be solved) in seismology to retrieve earthquake kinematic parameters such as the **seismic moment**, **slip distribution**, and **rupture velocity**

Rupture Area Portion of the fault surface that slips during an earthquake

Rupture Velocity Velocity at which the rupture front propagates over the fault surface during an earthquake; it may be very irregular spatially

Slip Distribution Pattern that describes the earthquake rupture in terms of timing and amount of slip along the fault surface

Earthquake Size

Moment Magnitude Measure of earthquake magnitude computed using the **seismic moment**, M_0 , estimated from various methods ($M_w = \frac{2}{3} \log(M_0) - 10.73$, with M_0 in dyne * cm)

Rigidity (or Shear Modulus) Shear **stress** to shear **strain** ratio of a material; this coefficient is used to measure the stiffness of a material and describe its response to applied shear stress; in **subduction zone** environments, the rigidity varies with depth, with lower values in the shallow sedimentary part of the **megathrust**

Seismic Moment A scalar measure of the earthquake size based on the rupture area (A), the average slip ($\bar{D}$) on the fault surface, and the rigidity (μ) of the medium surrounding the fault; the relationship among these parameters is $M_0 = A * \bar{D} * \mu$

Seismic Cycle

Coseismic Phase The interval in which the accumulated **stress** on the fault is suddenly released causing an earthquake

Interseismic Coupling The capability of accumulating stress on a fault in the period that separates two large seismic events (**interseismic phase**); strongly coupled (or locked) faults accumulate stress at the rate at which the plates are moving relative to each other and, in principle, are more prone to host large earthquakes

Postseismic Phase The interval in which additional slip or viscoelastic relaxation occurs after the mainshock as effect of the stress and strain redistribution

Seismic Cycle The three phases involving the process of stress and **strain** buildup (**interseismic**), release (**coseismic**), and redistribution (**postseismic**) that occurs along the fault surface before, during, and after an earthquake, respectively; each phase occurs with a very different time scale (many years for interseismic, seconds for coseismic, from minutes to months for postseismic)

Strain Is the deformation of a rock mass subjected to tectonic stresses

Stress Is the force per unit area applied on a surface (e.g., a fault surface) within a volume

Subduction

Megathrust An extremely large thrust fault, formed along a **subduction zone** plate boundary, hosting very large earthquakes

Subduction Zone Convergence zone of two tectonic plates where one plate moves beneath the other; the contact surface between the two convergent plates is termed the **subduction interface or megathrust**; subduction zones are the most seismically active environments on Earth and where most of the largest earthquakes occur

Trench The deep bathymetric trough position where the subducting plate begins to descend beneath the overriding plate

Wedge The shallow portion of the overriding plate characterized by the accumulation of sedimentary material due to the collision with the subducting plate

Tsunami

Run-up The maximum topographic elevation reached by the tsunami during the inundation

Tsunami Amplitude The difference, computed on a tsunami waveform, between the peak of the tsunami wave and the undisturbed sea-level value

Tsunami Height The difference, computed on a tsunami waveform, between successive crest and trough values of a tsunami wave

Tsunami Period The time between two successive tsunami wave extrema (generally considering the crest of the wave) that is the amount of time taken by a tsunami wave to complete a cycle

Tsunami Wavelength The distance between two consecutive tsunami waves (generally considering the crest of the wave)

Definition of the Subject

The decade 2004–2013 was characterized by an unusually large rate (1.7 per year) of great earthquakes ($M_W \geq 8$) (Lay 2015). The majority occurred on subduction zone megathrusts. Significant tsunamis were generated by most of them, as well as by several major (M_W 7.0–7.9) earthquakes at very shallow depth on megathrusts, or by great intraplate normal faulting events near subduction zones.

Two of the great events, the 2004 Sumatra-Andaman M_W 9.2 and the 2011 Tohoku M_W 9.0 earthquakes are, respectively, the third and the fourth largest earthquakes seismologically recorded (since 1900) and the largest events of the last 50 years (http://earthquake.usgs.gov/earthquakes/world/10_largest_world.php). These events were larger than had been anticipated for their source regions and produced two of the most damaging tsunamis in modern times. It is now widely accepted that great tsunamigenic earthquakes may occur in most subduction zones worldwide, although very infrequently in some regions (e.g., Stein and Okal 2011; Kagan and Jackson 2013).

The occurrence of many important seismic events during a time of greatly expanded global seismological, geodetic, and tsunami recording systems has provided the scientific community with an unprecedented set of geophysical observations. Data availability has prompted the development of new data inversion techniques for reconstruction and understanding of the space-time fault sliding process during the earthquake,

which, in turn, provides a better understanding of the resulting tsunamigenic potential.

We address most of the major and great tsunamigenic earthquakes that occurred over the decade. The objective of this entry is to discuss the general features of source models – as retrieved by inversion of seismic, geodetic, and tsunami data – and of the observed tsunamis. We highlight the kinematic complexity of the rupture processes in subduction zones and the relation between fault slip distribution, bathymetry, and coastal features on one hand and the observed tsunami impacts on the other hand.

Introduction

The seismological instrumental era begun in the early twentieth century (http://earthquake.usgs.gov/learn/topics/seismology/history/history_seis.php), as did systematic production of global seismic event catalogs (<http://www.isc.ac.uk/iscgem/>). Self-recording tide gauges (<http://noc.ac.uk/science-technology/climate-sea-level/sea-level/tides/tide-gauges/tide-gauge-instrumentation>) have been recording tsunamis since the mid-nineteenth century, whereas the tide-gauge catalog seems to be globally complete since the mid-twentieth century (Geist and Parsons 2011; <https://www.ngdc.noaa.gov/hazard/tide.shtml>).

The largest megathrust earthquakes seismologically recorded prior to the decade 2004–2013 were the 1952 M_W 9.0 Kamchatka, 1964 M_W 9.2 Alaska, and 1960 M_W 9.5 Chile earthquakes (Kanamori 1977; http://earthquake.usgs.gov/earthquakes/world/10_largest_world.php), all of which generated transoceanic tsunamis in the Pacific. The cumulative global seismic moment release of those three events exceeds that of all great events of the recent decade, but the rate of great earthquakes was higher from 2004 to 2013 than previously observed (Lay 2015). Most of these great earthquakes occurred near the Pacific and Indian Plate boundary margins where previous great earthquakes have occurred, but ~40 % ruptured within lithospheric plates rather than directly on plate boundaries, and the location of

the latter events could not be directly anticipated based on plate motions. Even for the plate boundary events, some of the ruptures had surprising spatial extent and features of their slip distribution, as well as nature of the tsunamis they generated.

The overall size of the 2004 M_W 9.2 Sumatra-Andaman 2004 megathrust earthquake had not been anticipated for the region along on the Sunda Trench from Northern Sumatra to the Andaman Islands. This was due in part to the lack of knowledge of its subsequently discovered middle-ages predecessors (Sieh et al. 2015 and references therein) and in part because of widely accepted assumptions about the seismic potential of the highly oblique convergence of oceanic lithosphere along this plate boundary. The rupture length (>1300 km) and duration (~ 8 min) were unprecedented. This earthquake rupture spread across the rupture zones of several prior major-to-great events (1847, 1881, 1941), without being constrained by any lateral segmentation of the megathrust. The very long rupture duration necessitated improvements of seismological inversion methods for finite source models (see, e.g., Shearer and Bürgmann 2010). For example, the first reliable estimate of moment magnitude was made long after the earthquake (e.g., Park et al. 2005), and it took several months to converge on reliable finite-fault models from seismology (e.g., Ammon et al. 2005) and even longer for analysis of the measurements from campaign geodesy (e.g., Vigny et al. 2005).

This earthquake generated a huge tsunami that expanded throughout the Indian Ocean and into the Pacific and Atlantic Oceans (Titov et al. 2005). The catastrophic loss of life ($\sim 228,000$ people) triggered global efforts aimed at providing tsunami warning systems for more regions than the Pacific. This includes the Caribbean and Mediterranean, where tsunamigenic events have a low occurrence rate and the hazard from infrequent catastrophic events was underappreciated prior to the 2004 Sumatra event.

The December 2004 event was the first in a sequence of earthquakes and tsunamis to strike along the Sunda Trench from 2004 to 2010 that we describe in some detail in this entry and which

included not only “ordinary” megathrust earthquakes but also two “tsunami earthquakes.” Within the Indo-Australian plate seaward of the 2004 rupture, the largest ever recorded strike-slip earthquake (M_W 8.7) also occurred in 2012, generating a small tsunami.

Similar to the 2004 Sumatra-Andaman earthquake, the Tohoku 2011 M_W 9.0 earthquake was larger than had been anticipated in the region, mostly because of incomplete information about prior earthquakes and incorrect assumptions concerning fault segmentation and the maximum possible magnitude (Stein and Okal 2011). The rupture length of significant slip spanned only ~ 350 km along strike, much shorter than the 2004 Sumatra earthquake. Both events have quite extreme values compared to the expected mean value for earthquake scaling laws (e.g., Strasser et al. 2010). How large the variability of rupture size can be is also appreciated considering that the 2010 Maule earthquake (M_W 8.8, corresponding to a two to three smaller seismic moment) ruptured a fault length of ~ 500 km.

The 2011 Tohoku event involved a very large amount ($> \sim 50$ m) of slip reaching all the way to the trench, which strongly increased the tsunami excitation. It is commonly thought that the shallow sedimentary wedge cannot accumulate significant elastic strain and the shallow fault friction should be slip velocity strengthening, favoring aseismic creep rather than stick-slip earthquake failure (e.g., Wang and Kinoshita 2013), so the large coseismic slip at shallow depth was surprising. Dynamic rupture processes appear to have played a role in this large slip at shallow depth.

An important context for anticipating future strong earthquakes is the identification of seismic gaps (e.g., Nishenko 1991), i.e., regions where large documented earthquakes occurred in the past and the time since the last large rupture is a significant fraction of the average recurrence interval (which can be inferred from frequency of past events or from the strain accumulation rate). Based on the elastic rebound mechanism of stress/strain accumulation during the interseismic period that results from frictional resistance to motion between two moving plates

and the sudden stress/strain release during subsequent earthquake sliding, the earthquake occurrence probability on a fault may (simplistically) tend to increase with time since the last earthquake. Over the past several decades, most great subduction zone megathrust earthquakes have in fact occurred or initiated within previously identified seismic gaps or regions that at least had unknown prior large event history (e.g., Lay 2015). This may reflect the relative simplicity of the seismic cycle involved in subduction zone plate convergence. However, the timing of future events is not well-constrained, and seismic gaps are often not robustly identified due to limited knowledge of prior events. Additionally, the rupture areas have not always conformed to identified gaps, likely due to the nonlinear cascade involved in large earthquake ruptures; earthquake ruptures are affected by local scale frictional and geometrical heterogeneity, and dynamic rupture propagation may occasionally overcome “barriers” of relatively low stress such as slip velocity-strengthening regions, as shown by numerical models (e.g., Kaneko et al. 2010). Some events only partially ruptured seismic gaps (e.g., the 2001 Peru (M_W 8.4) event ruptured only 60 % of the 1868 rupture zone in southeastern Peru, and the 2014 Iquique, Chile (M_W 8.0), event ruptured only 20 % of the 1877 rupture zone in northern Chile); others ruptured the megathrust well beyond identified gaps (e.g., the Maule M_W 8.8 and the 2004 Sumatra M_W 9.2 earthquakes); yet others rupture the whole megathrust, as the 2011 Tohoku M_W 9.0 earthquake did, breaking across inferred regions of lateral down-dip-up-dip segmentation based on previous earthquake history (e.g., Stein et al. 2012). Together with the partial re-rupture in the Mentawai gap of the 1797 Sumatra rupture zone by the large events in 2007, these provide examples of variable megathrust failure indicating that rupture zone-based segmentation is not a reliable guide to future earthquake size. The resulting variability of large earthquake ruptures has prompted questions as to the statistical value of the gap model (e.g., Jackson and Kagan 2011).

Subduction zones also occasionally host the abovementioned “tsunami earthquakes,” addressed

elsewhere in this encyclopedia (Polet and Kanamori 2009). These are generally major ($M_W < 8$) and very shallow earthquakes that generate a tsunami larger than expected for their magnitude. Observationally, they feature large slow slip and relatively long duration and weak ground motion. For this reason they rarely trigger self-evacuation of coastal populations based on the recognition of strong shaking. Their rupture process is difficult to model and their frequency of occurrence is difficult to assess; thus, they are often overlooked in tsunami hazard assessments (e.g., González et al. 2009; UNISDR 2015). We here discuss two tsunami earthquakes that occurred in the recent decade along the Sunda Trench, the 2006 Java and 2010 Mentawai earthquakes. Another case challenging the common expectation of aseismic or slow-slip behavior of the accretionary wedge that we discuss is the Haida Gwaii (2012) earthquake.

Other earthquakes that occurred in the decade 2004–2013 demonstrated additional complexity of the earthquake rupture process and tsunami generation. For example, earthquake doublets demonstrate triggering interactions between two regional fault systems and increased overall tsunami hazard, while events located in complicated tectonic environments, such as deformed slabs or junctions of three plates, also illustrate the difficulty of anticipating tsunami excitation. Understanding the complexity of ruptures is complicated by the nonuniqueness of inversion for the earthquake slip distribution and challenging model uncertainty quantification, and this enhances the value of analyzing seismic, geodetic, and tsunami data collectively to better constrain the solutions.

The next section gives a brief discussion of some important concepts regarding tsunami generation by earthquakes and their impact on local and distant coastlines (“Tsunamis and the Seismic Source”). Then we remark on inversion for the earthquake source by seismic, geodetic, and tsunami data (“Inversion for the Seismic Source”). The rest of the entry is a summary of selected tsunamigenic major-to-great earthquakes that occurred in the decade 2004–2013. It begins with “Megathrust Events,” including those in

“The 2004–2010 Sunda Trench Sequence” (“Sumatra-Andaman (2004),” “Nias (2005),” “Bengkulu (2007),” “Peru (2007),” “Maule (2010),” “Tohoku (2011),” and “Santa Cruz Islands (2013).” Then we describe the “Great Subduction Earthquakes Doublets” involving paired outer-rise and megathrust earthquakes: the “Kurils (2006, 2007)” and the “Samoa (2009)” doublets. “Strike-slip Events” section follows, describing the “Sumatra (2012)” earthquakes. Large “Tsunami Earthquakes” section deals with “Java (2006)” and “Mentawai (2010)” events. Some “Special Cases” of tsunamigenic earthquakes complete this overview (“Solomon Islands (2007),” “Haida Gwaii (2012)”).

Tsunamis and the Seismic Source

Tsunamis are water gravity waves resulting from seafloor deformation induced by submarine or coastal earthquakes (tsunamigenic earthquakes). Less frequently, tsunamis are generated by submarine landslides, volcanic eruptions, atmospheric phenomena, or meteorite impacts. The majority of tsunamigenic earthquakes occur along subduction zones as “ordinary” interplate thrust earthquakes (rupturing on the shallow dipping plate boundary interface); less frequently they show a “tsunami earthquake” character. Significant tsunamis are sometimes generated by intraplate normal faulting earthquakes (e.g., outer-rise events). Also oceanic strike-slip earthquakes can be tsunamigenic in the presence of steep bathymetric slopes. Magnitude and features of earthquake-generated tsunamis depend on several factors.

Great megathrust earthquakes located for example on the Pacific “ring of fire” commonly generate tsunamis that propagate to long distances. To first order, the tsunami amplitudes and mean run-up recorded at long distances scale with earthquake magnitude (Okal 1988; Geist 2009); nevertheless, tsunami impact can be heterogeneous depending on source orientation and bathymetry features. The tsunami generation process by complex earthquake slip distributions has been discussed within this

encyclopedia (e.g., Geist and Oglesby 2014 and references therein). A major control factor is the amount and distribution of vertical seafloor coseismic displacement, particularly in the near field, which depends not only on the earthquake magnitude but also on the focal depth and mechanism of the earthquake and on the slip distribution. The complex relation between slip distribution and tsunami features in the near field may depend also on the relative position and orientation of the source and the coastline, the water depth above the source zone, the steepness of the initial tsunami field, and the periods of the tsunami compared to the size of the continental shelf and to the features and the complexity of the coastline where the inundation occurs.

Inversion for the Seismic Source

The earthquake source is commonly investigated by inversions of geophysical data in order to retrieve the kinematic parameters of the earthquake rupture, such as the space-time slip distribution, the rise time for slip at a point on the fault, and the rupture expansion velocity. Generally, some assumptions are needed to limit nonuniqueness of the problem, such as a priori information on the fault geometry or positivity of the slip vector, and regularization of the solution (e.g., Das 2011).

Inversions of tsunami data for the earthquake kinematic parameters (Satake 2009, this encyclopedia, and references therein) are usually performed by linear least-squares method or global search techniques (e.g., Piatanesi and Lorito 2007). In case of a tsunamigenic event whose causative source is thought to be partially or totally non-seismic (e.g., a submarine landslide) or if fault geometry is not enough constrained, tsunami travel-time inversion is sometimes used to determine the location and extent of the tsunami source (e.g., Lay et al. 2005a; Neetu et al. 2005; Ishii et al., 2012), or tsunami and other geophysical data are inverted to retrieve the initial sea surface displacement instead of the kinematic earthquake source

parameters (e.g., Hossen et al. 2015 and references therein).

More generally, different inversion techniques for the seismic source can be used, depending on both the kind and complexity of the problem that has to be solved, such as linear least-squares method (e.g., Hartzell and Heaton 1983; Satake 2009; Yue and Lay 2011), Bayesian inversion (e.g., Minson et al. 2013), or global search methods (e.g., simulated annealing: Rothman 1986; Piatanesi and Lorito 2007; Romano et al. 2014).

The data used for inversion depend on the available instruments that recorded the earthquake. For example, global seismic data are regularly used for investigating all moderate to great earthquakes; strong-motion and geodetic data (e.g., GPS, land leveling) are very useful if the seismic source is relatively close to an instrumental network. Recently, differential interferograms from InSAR, potentially giving a wider image of the inland displacement, and high-rate GPS data, complementing strong-motion seismic data at lower frequencies, are being successfully used in inversions. Tsunami data measured by tide gauges, bottom pressure sensors, and more recently GPS buoys can be used if the earthquake generated a tsunami large enough to be instrumentally recorded (see Satake 2009; this encyclopedia); occasionally, also satellite altimetry has detected the tsunami waves in the open ocean.

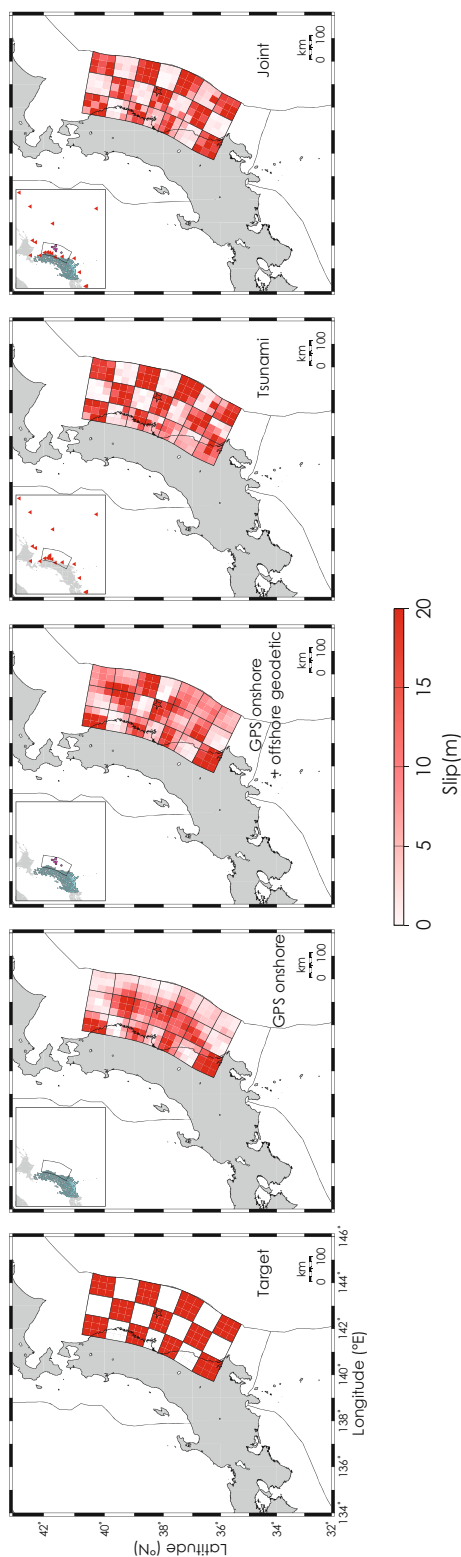
The accuracy of the retrieved seismic rupture image depends on the data used, and each observable has its specific resolving power for different rupture features (e.g., Lay 2015). For example, seismic waves are quite sensitive to faulting orientation and the time function of the source, while tsunami waves are primarily sensitive to total fault slip and long-wavelength slip features. Teleseismic data also provide reliable seismic moment estimation, but, compared to tsunami data, they may have a lower resolving power for the shallow offshore slip; static geodetic data can constrain the seismic moment, but they only resolve local slip. An increasingly used strategy is to perform joint inversions of different kinds of data, in order to exploit their possibly complementary resolving power and/or sensitivity to different

aspects of the seismic rupture (e.g., Romano et al. 2012; Yue et al. 2015). Relative weights to be assigned to the different datasets are difficult to assess; synthetic checkerboard-like slip models are commonly employed for assessing the resolution offered by individual datasets and for calibrating their relative weights for the joint inversion (Fig. 1). However, there are still several challenges to be faced: models that satisfactorily match one data type may fail to match others, uncertainty in model parameterization, and noncomparable accuracy in some cases of the forward modeling techniques. Also, the wide range of fault deformation time scales that is now being recognized and the intrinsic variations in sensitivity to time scales of the source deformation need to be taken into account; for example, deformation (or even slumps) too slow for seismic wave excitation may still excite tsunami, while less tsunamigenic deep ruptures may provoke strong ground motion; so it may be difficult to force consistency with a single representation of the source. Robust assessment of the earthquake kinematic model uncertainties remains an open research problem.

Megathrust Events

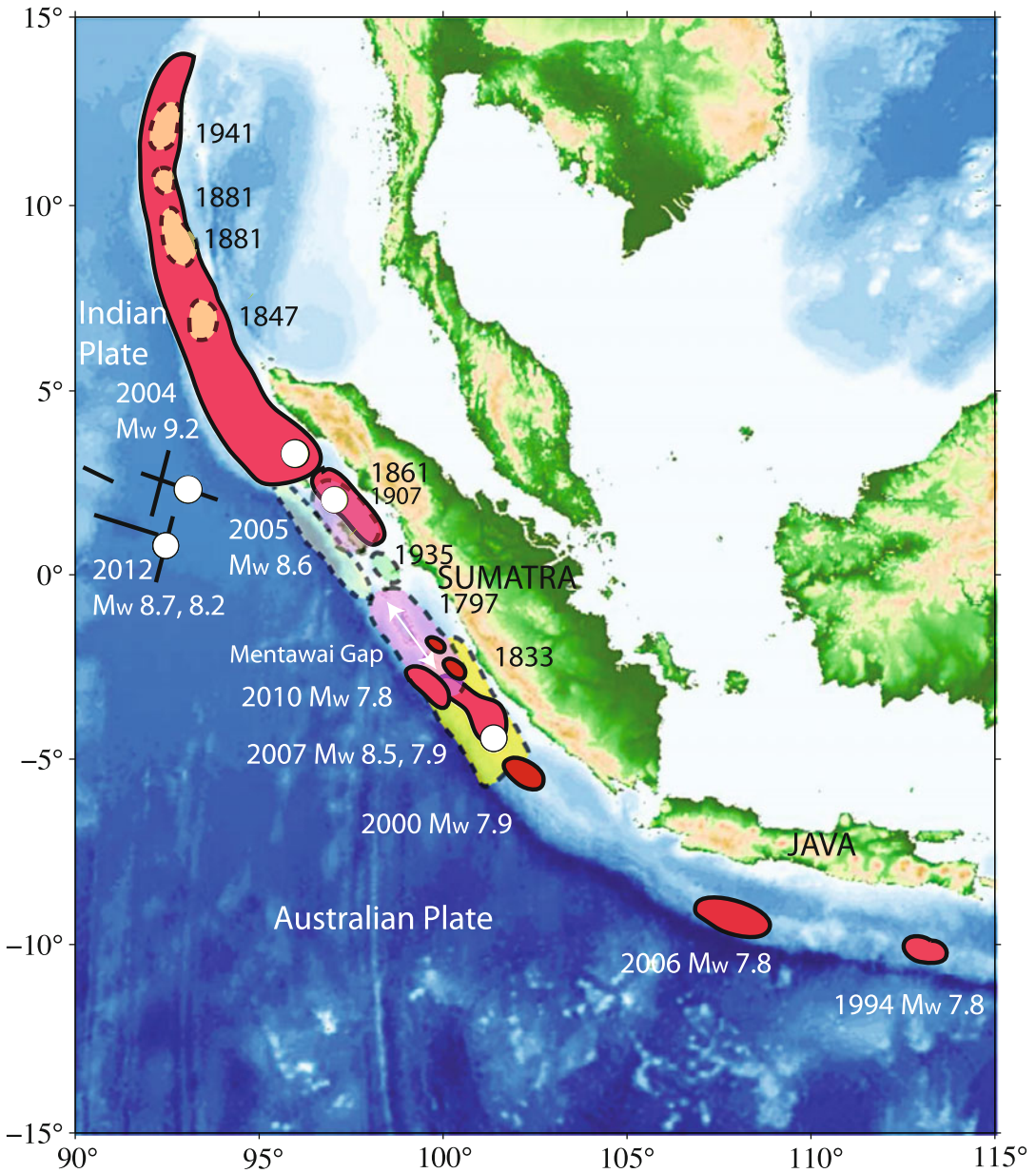
The 2004–2010 Sunda Trench Sequence

The Sunda Trench in the Indian Ocean (Fig. 2) marks the partitioned oblique convergence of the Indo-Australian Plate under the Burma silver microplate along the Andaman and Nicobar Islands and under the Sunda Plate along Indonesia. The Sunda Trench forms a curved arc trending roughly NW-SE, extending from the Andaman Islands, along Sumatra and Java, to the Lesser Sunda Islands. Between 13°N and 7°S, the trench-parallel component of the relative plate motion is accommodated by the Andaman-Nicobar back-arc system and the Great Sumatran fault. From 2004 to 2010 the Sunda megathrust ruptured over much of its length in a sequence of tsunamigenic earthquakes, including three great megathrust earthquakes in 2004 (M_w 9.2), 2005 (M_w 8.6), and 2007 (M_w 8.4) and four M_w 7.6–7.9 earthquakes (2006, 2007, 2009, 2010),



Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 1 Checkerboard test for individual and joint inversions. In a checkerboard test, a specified pattern of final fault slip is prescribed, and corresponding predictions are made for geodetic, seismic, and tsunami waves generated by that pattern. The *left* panel shows a checkerboard pattern of unit and zero slip over a fault offshore of Japan. The predictions (usually with some noise injected) are made for either a dynamic rupture expansion model or a static

model, at locations corresponding to actual data observations for each wave type. The computed signals are then separately or jointly inverted to try to retrieve the input model, with the intrinsic limitations of each data type and recording configuration limiting the recovery of the target checkerboard pattern. In principle, combining GPS and offshore geodetic and tsunami data allows the best reconstruction of the complete checkerboard target model; however, this kind of test does not account for data modeling uncertainties (From Romano et al. 2012)



Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 2 Tectonic plates and earthquakes along the Sunda Trench (From

Lay 2015) (Reprinted from Earth and Planetary Science Letters, Vol. 409, Lay T., The surge of great earthquakes from 2004 to 2014, pp.133-146, Copyright 2015, with permission from Elsevier)

two of which are classified as “tsunami earthquakes.” Additionally, two great strike-slip earthquakes (M_w 8.7, 8.2) occurred within the Indo-Australian Plate seaward of the trench in 2012. A comparable spatiotemporal clustering of great

earthquakes occurred along the Alaska-Aleutians arc from 1938 to 1965 (Lay 2015) including the 1964 $M_w \sim 9.2$ Alaska earthquake and four great earthquakes in 1938, 1946 (a tsunami earthquake), 1957, and 1965.

The M_W 9.2 Sumatra-Andaman earthquake is the first giant earthquake to be observed by the modern global network of broadband seismometers and GPS campaign networks (Subarya et al. 2006; Ammon et al. 2010). No prior event had been historically recorded along northwestern Sumatra where the 2004 earthquake had the largest slip.

The Sumatra-Andaman earthquake caused the deadliest tsunami of modern times. Together with the tsunami generated by the 21 July 365 Crete M_8+ earthquake, it is likely one of the two deadliest tsunamis ever. The coast of northern Sumatra suffered the greatest losses, but all margins of the Bay of Bengal including Thailand, Sri Lanka, India, and Myanmar were hit within a few hours; 5–10 h later the Maldives and Somalia experienced damaging tsunami. Very little warning was given even far from the event, as there was no operational tsunami warning system in the Indian Ocean. In response to this tragic event, IOC/UNESCO received a mandate from the international community to coordinate the establishment of regional tsunami early warning systems globally, which are now largely operational (<http://www.ioc-tsunami.org>).

The adjacent segment of the Sunda Trench immediately southeast of the 2004 event ruptured 3 months later in the March 2005 M_W 8.6 Nias earthquake. The rupture zone of the 2005 event overlaps the deeper part of the estimated source zone of a great 1861 earthquake (Natawidjaja et al. 2006). The shallow portion of the 1861 rupture may have failed in the 1907 tsunami earthquake (Kanamori et al. 2010). Due to the characteristics and location of the Nias earthquake slip distribution, discussed in the next section, the ensuing tsunami was weak relative to the earthquake magnitude (Geist et al. 2006), demonstrating how difficult it can be to forecast tsunami impact based only on hypocenter and magnitude of an earthquake.

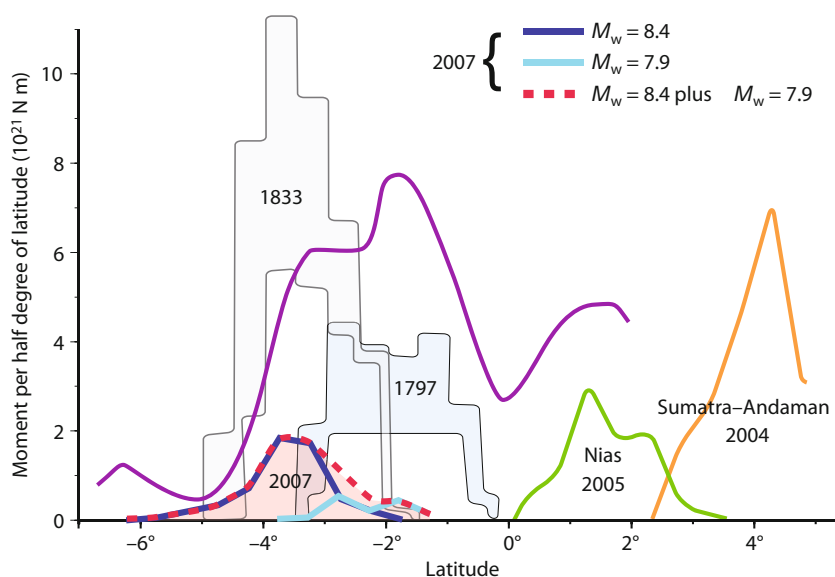
Southeast of the 2005 rupture zone, a mostly locked section of the megathrust extends about 200 km along the Mentawai Islands approximately to Enggano Island, where the last great earthquake occurred in 1797 (Philibosian et al. 2014). The subsequent 2006–2010 sequence straddles this region.

A major M_W 7.8 tsunami earthquake ruptured far offshore of Java in 2006, producing a damaging tsunami; then in September 2007, the great M_W 8.4 Bengkulu earthquake struck in the region of a great 1833 earthquake, which overlapped the southeastern portion of the 1797 event. This event was followed 12 h 39 min later by an M_W 7.9 aftershock, which ruptured the down-dip portion of the overlapping zones of the 1797 and 1833 ruptures. The two 2007 events generated modest tsunamis. When a large M_W 7.6 81 km deep intraslab earthquake occurred in September 2009 below Padang, the city was heavily damaged by the strong shaking, and people fled to high ground in fear of a possible tsunami. Due to the focal depth and location of the 2009 earthquake, the tsunami was very small and only instrumentally recorded. The up-dip portion of the overlapping zones of the 1797 and 1833 ruptures was then struck by a tsunami earthquake in 2010, located seaward of the Pagai Islands and up-dip from the two 2007 earthquake source zones. There is a ~400 km long section of the Sunda Trench between the 2005 and 2007 rupture zones, the “Mentawai gap” (e.g., McCloskey et al. 2010), with remaining potential for large-to-great megathrust earthquakes and tsunamis according to the estimated moment deficit (Konca et al. 2008; Fig. 3).

Sumatra-Andaman (2004)

The giant Sumatra-Andaman megathrust earthquake (M_W 9.2) nucleated ~80 km off the northwestern coast of Sumatra and propagated northward for ~1300 km (Fig. 4), causing a devastating tsunami in the Indian Ocean.

Different teams (Choi et al. 2006 and references therein) conducted post-tsunami surveys on Indian Ocean coasts, collecting more than 1500 run-up measurements. The highest run-up occurred in the Aceh region (northern Sumatra), near the maximum slip featured by most of the finite-fault models. The tsunami penetrated more than 2 km inland, and run-up values were as high as ~50 m at some places (s). Aceh province had the highest death toll (160,000+ victims of a total of ~228,000). The impact was extensive in many countries: Indonesia (average run-up ~10 m),



Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 3 The purple line represents the accumulated moment deficit since the last great rupture in the Mentawai gap (From Konca et al. 2008) (Reprinted by permission from Macmillan Publishers Ltd: [Nature], Konca A O, Avouac J P,

Sladen A, Meltzner A J, Sieh K, Fang P, Li Z, Galetzka J, Genrich J, Chlieh M, Natawidjaja D H, Bock Y, Fielding E J, Ji C, Helmberger D V, Partial rupture of a locked patch of the Sumatra megathrust during the 2007 earthquake sequence, *Nature*, 456:631–35, doi:10.1038/nature07572, Copyright 2008)

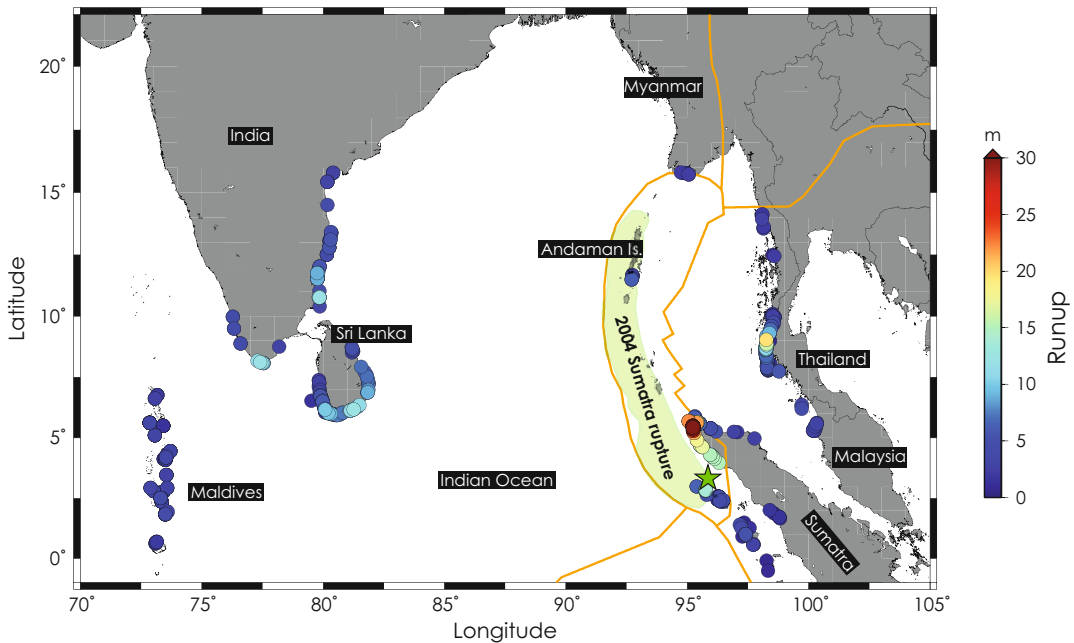
Thailand (average run-up ~ 5 m), Sri Lanka (average run-up ~ 5 m), and India (average run-up ~ 4 m), and significant waves reached across the Indian ocean striking, for example, the coasts of Somalia (run-up ~ 10 m in some places), Bangladesh, Malaysia, Maldives, and Seychelles Islands.

The 2004 tsunami propagated well beyond the Indian Ocean (Fig. 5), and for the first time a tsunami was measured by worldwide tide gauges. Numerical simulations revealed how bathymetric waveguides trapped and directed tsunami energy; for example, the Southwest Indian and Mid-Atlantic ridges guided the tsunami toward the Atlantic Ocean, whereas Southeast Indian and Pacific-Antarctic ridges guided it toward the Pacific Ocean (Titov et al. 2005).

The tsunami waves at tide gauges located in the far field (e.g., Callao in Peru, $\sim 19,000$ km from the epicenter, or Halifax in Nova Scotia, $\sim 24,000$ km) were sometimes higher than those measured in the near field (e.g., Cocos Island), due to amplification effect during propagation on the

continental shelf, enhanced by local resonances (e.g., Thomson et al. 2007). Commonly, in the far-field tsunami signals, the maximum wave heights arrive several hours after the leading wave due to multiple reflections and coastal resonances (see also discussion of the 2006 M_W 8.3 Kuril earthquake); this is a very important factor when evaluating an “all clear” announcement after issuing a tsunami alert message.

Many studies were performed as inversion techniques were progressively updated to handle the long rupture duration (e.g., Ammon et al. 2005; Gahalaut et al. 2006; Banerjee et al. 2007; Chlieh et al. 2007; Piatanesi and Lorito 2007; Rhie et al. 2007; and many others). These studies used different kinds of geophysical data (teleseismic, tsunami, GPS, satellite altimetry) and different inversion techniques, which each offer variable resolution of the source process (see discussions in Menke et al. 2006; Shearer and Bürgmann 2010; Poisson et al. 2011; Lay 2015). Notably, sea-level anomalies up to ~ 80 cm were measured in the open



Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 4 Extent of the 2004 Sumatra-Andaman rupture and run-up distribution in the Indian Ocean

Indian ocean by multiple-satellite altimetry passes of Jason-1 and TOPEX/Poseidon satellites and were used for the first time in source inversions (Fujii and Satake 2007; Pietrzak et al. 2007; Hoechner et al. 2008; Lorito et al. 2010).

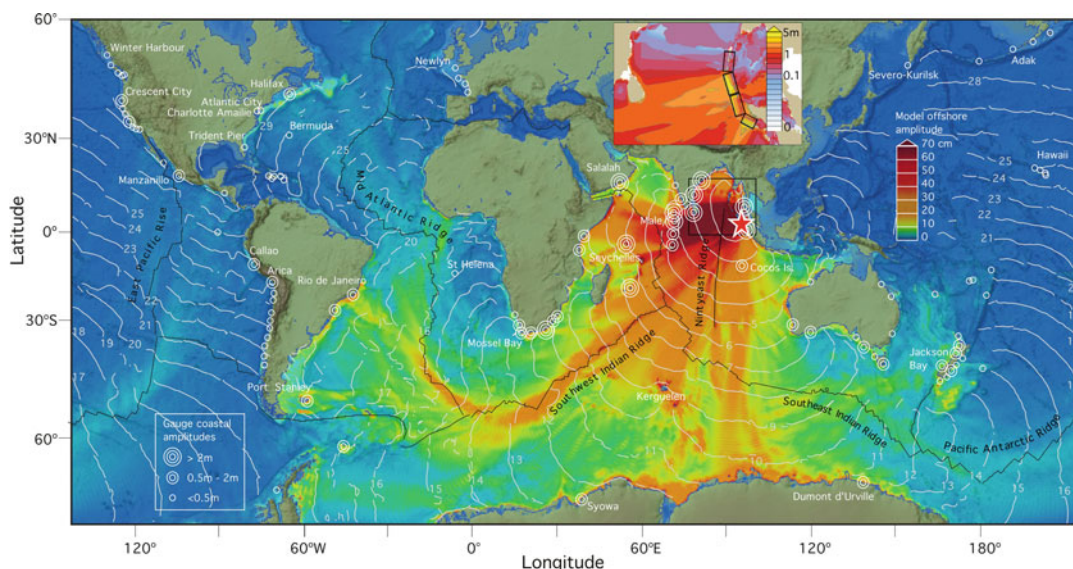
The source models are still not fully in agreement about the total seismic moment (estimates of M_W range from 9.2 to 9.3) or the slip location and amount (estimates of maximum slip range from 11 to 35 m, although this is usually a poorly resolved value for slip inversions). While the models also include a wide range of rupture velocities (from ~ 1 to ~ 4 km/s), several models indicate that the rupture progressively slowed down northward, and slow slip in the northernmost segment may have occurred, as suggested by geodetic observations (Bilham 2005; Lay et al. 2005b). Guided by different analyses, others argue against this possibility (Ishii et al. 2007; Shearer and Bürgmann 2010). The very large local run-up values remain underpredicted by most of the published slip distributions. Coseismic activation of a splay fault on the western edge of Banda Aceh has been proposed as a possible explanation (Plafker et al. 2006), and

seismic evidence for disrupted oceanic crust in the epicentral region has been reported (Singh et al. 2008).

Nias (2005)

Three months later, the great megathrust Nias earthquake (28 March 2005, M_W 8.6) occurred between the Nias and Simeulue Islands, ~ 200 km southeast from the great 2004 earthquake epicenter. Several studies suggest that this event was triggered by the stress increase resulting from the adjacent 2004 Sumatra earthquake (Nalbant et al. 2005), with a possible poroelastic component (Hughes et al. 2010).

The 2005 tsunami mainly struck the coasts of Simeulue Island (run-up up to ~ 4 m; Borrero et al. 2012), Nias, Banyak, and Batu Islands above the rupture area. Tsunami waves up to ~ 1.2 m (peak to trough) were recorded by the Sibolga tide gauge, in the near field of the source, whereas only centimetric waves were measured in the far field by several tide gauges located along the coasts of Oman, Sri Lanka, Mauritius, Maldives, and Seychelles Islands.



Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 5 Global reach of the 2004 tsunami indicated by peak wave height predictions (Titov et al. 2005 (Reprinted with permission from AAAS). (Readers may view, browse, and/or download material for temporary copying purposes only,

provided these uses are for noncommercial personal purposes. Except as provided by law, this material may not be further reproduced, distributed, transmitted, modified, adapted, performed, displayed, published, or sold in whole or in part, without prior written permission from the publisher)

Source models from teleseismic, GPS, and coral displacement inversions (e.g., Briggs et al. 2006; Hsu et al. 2006, 2011; Konca et al. 2007) inferred two patches of slip (up to ~ 12 m) mainly located under Nias and southern Simeulue Islands (Fig. 6). In addition, analysis of postseismic deformation highlighted significant aseismic afterslip, equivalent to an M_W 8.2 event, during the 11 months after the mainshock (Hsu et al. 2006), both up-dip and down-dip of the main coseismic rupture area (Fig. 6).

Despite the great magnitude, the tsunami was moderate, due to the depth of the slip distribution and the vertical displacement field lying almost entirely under the islands or relatively shallow water (e.g., Geist et al. 2006). This is noteworthy compared to the tsunamigenic excitation by earthquakes with smaller magnitude (e.g., tsunami earthquakes or the shallow 2007 Solomon Island rupture). Maximum run-up was observed along the southeastern tip of Simeulue Island located facing the northern patch of the inferred slip distribution.

Bengkulu (2007)

A great M_W 8.4 interplate earthquake occurred on 12 September 2007, with an epicenter located ~ 130 km offshore from the city of Bengkulu. It was followed 12 hr 39 min later by an M_W 7.9 aftershock to the northwest. Source models of the M_W 8.4 earthquake obtained by inverting, separately or jointly, geophysical datasets including teleseismic, GPS, coral uplifts, InSAR, and tsunami data (e.g., Lorito et al. 2008; Konca et al. 2008; Gusman et al. 2010) agree to first order on northwestward rupture expansion from the epicenter toward South Pagai Island, with maximum slip in the range of ~ 7 – 12 m close to a previously strongly coupled region of the interface.

The tsunami following the 2007 earthquake was moderate, likely because most of the rupture occurred relatively deep along the megathrust. The run-up measurements collected along the coast of the Bengkulu province are < 4 m (Borrero et al. 2009), and the closest tide gauge (Padang) measured a tsunami wave of ~ 1 m (peak to trough).

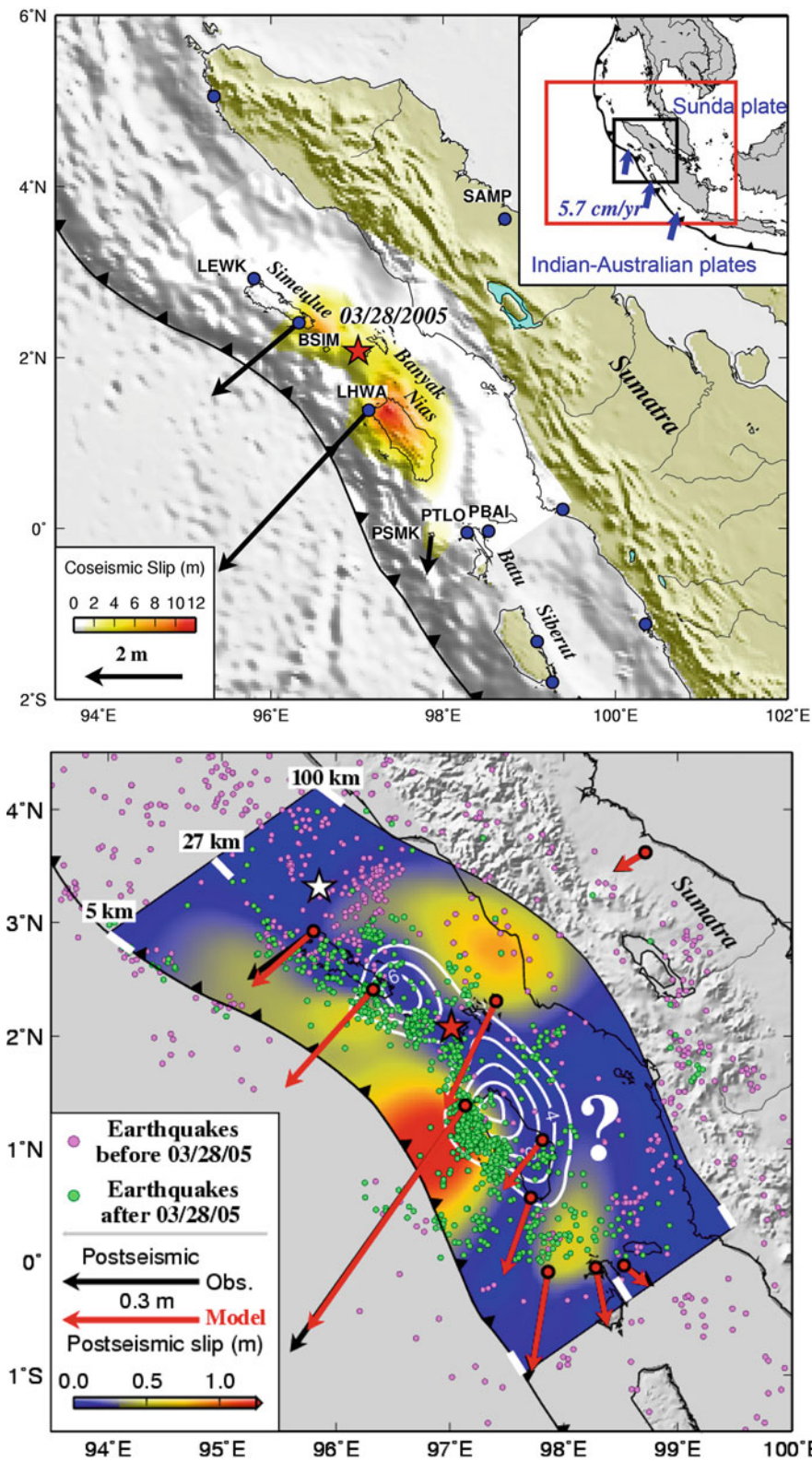


Fig. 6 (continued)

Peru (2007)

On 15 August 2007, an M_W 8.0 earthquake occurred offshore of central Peru ~ 60 km west from the city of Pisco. The hypocenter was located at a depth of ~ 40 km on the subduction interface between the Nazca and South American Plates. The earthquake generated a tsunami that struck the Peruvian coast from Lima ($\sim 12^\circ\text{S}$) to 50 km south of the Paracas Peninsula (14.5°S). The run-up measurements range from 1 to 10 m (Fritz et al. 2008). Three victims were reported in the area of maximum inundation at Lagunilla (Ioualalen et al. 2013). The moderate size of the rupture and its depth limited the regional tsunami excitation for this event. The 2007 Peru earthquake generated centimetric-to-decimetric tsunami amplitudes measured by tide gauges installed in French Polynesia, California, Mexico, Chile, New Zealand, Hawaii, and Japan.

Both teleseismic (Lay et al. 2010b) and InSAR (Sladen et al. 2010) data have been used to invert for finite-fault rupture models of the 2007 Peru earthquake. These models (Fig. 7) both have two patches of slip, one near the epicenter and the other located in the southwest off the Paracas Peninsula. The moment rate function (time history of moment release) features a first pulse within a few seconds after the rupture began and a second larger pulse ~ 60 s later (Lay et al. 2010b). This prolonged source process could conceivably be associated with a slow rupture velocity (< 1.5 km/s), which would be unusual for a rupture confined in the deeper part of the megathrust. There is no clear evidence from geodetic data for slow or aseismic slip on the megathrust in this region. Thus, it appears likely that the ~ 60 s hiatus in the seismic radiation is due to discrete failure of an initial smaller sub-event (M_W 7.8) that dynamically triggered the second larger sub-event (M_W 8.0) (Lay et al. 2010b). This type of compound failure also appears to have occurred

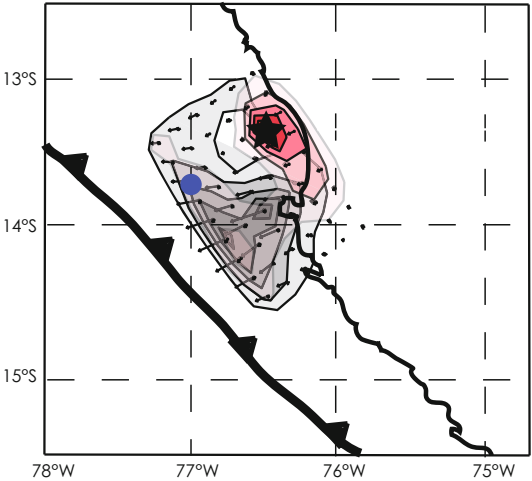
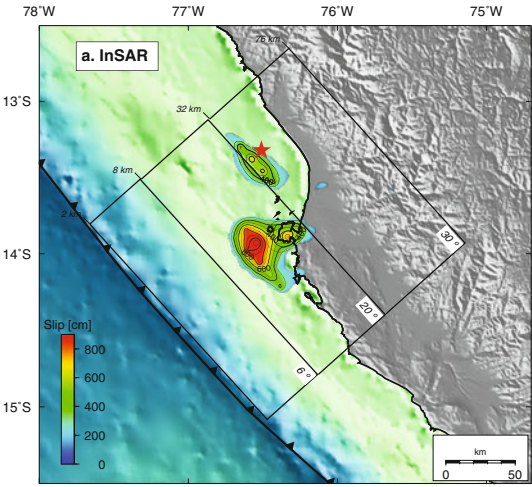
during the 2001 Peru M_W 8.4 earthquake (Lay et al. 2010b; Pritchard et al. 2007), which initiated with an M_W 7.5 rupture that appears to have dynamically triggered a larger rupture about 120 km away that grew into an M_W 8.4 event.

The 2007 Peru event, together with the adjacent 1940 (M_W 8.0) and 1974 (M_W 8.1) earthquakes, forms a patchwork collectively spanning the estimated source zone of the 1687 $M_{8.5+}$ earthquake (Biggs et al. 2009). Similarly, on the Colombia-Ecuador margin, the rupture zone corresponding to the 1906 earthquake (M_S 8.8, Kanamori and McNally 1982) re-ruptured in three smaller separate events (1942, M_S 7.9; 1958, M_S 7.7; 1979, M_W 8.2).

While the entire stretch of Peruvian coast from north of Lagunilla to Pisco runs broadside of the rupture zone, the largest run-ups were measured south of the Paracas Peninsula (Fig. 8). Indeed, the southern side of the peninsula was most affected by the tsunami, with run-ups from 5 to 7 m measured in Lagunilla and the maximum tsunami run-up of 10 m measured in Playa Yumaque. Lagunilla experienced the maximum inundation distance, as the tsunami penetrated 2 km inland on flat terrain. This pattern is likely an effect of the peninsula extending obliquely to the NNW above the maximum slip area, exposing its southernmost side to the leading wave, much more than the northernmost stretch of coast. On the north side of the peninsula, the measured run-ups were only 3 m on average. Additionally, the islands located west and north of the peninsula might have attenuated the tsunami impact north of Paracas Peninsula (Fritz et al. 2008; Ioualalen et al. 2013). Run-ups up to ~ 5 m were also measured south of Lagunilla, within the Bahia de la Independencia, which is located obliquely to the tsunami source. These southern surges may have been generated by edge waves, which are waves trapped by refraction to propagate along the coast.

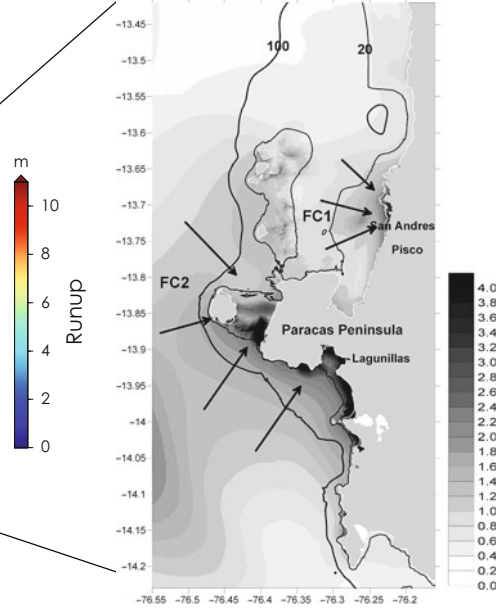
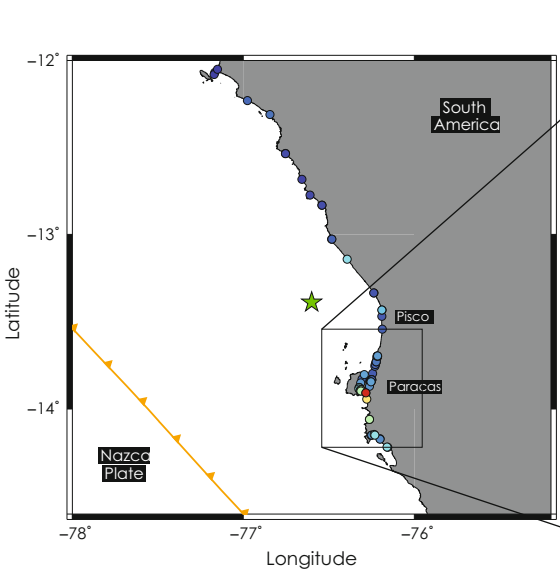
Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 6 Coseismic (*top* panel, from Hsu et al. 2011) and postseismic (*bottom* panel, from Hsu et al. 2006) (Reprinted with permission from AAAS) (Readers may view, browse, and/or download material for temporary copying purposes only,

provided these uses are for noncommercial personal purposes. Except as provided by law, this material may not be further reproduced, distributed, transmitted, modified, adapted, performed, displayed, published, or sold in whole or in part, without prior written permission from the publisher). panel, from Hsu et al. 2006) slip of the Nias 2005 earthquake



Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 7 Slip distributions estimated for the 2007 Peru earthquake (*left panel*

from Sladen et al. 2010; *right panel* from Lay et al. 2010b) peak slip for the first event (red tones) is 3.83m; for the second event (gray tones) is 9.14



Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 8 Run-up distribution (*left panel*) and simulated tsunami maximum wave heights for the 2007 Pisco earthquake (*right panel*

from Ioualalen et al. 2013) (Reprinted with kind permission from Springer Science+Business Media: Pure and Applied Geophysics, Tsunami Modeling to Validate Slip Models of the 2007 Mw8.0 Pisco Earthquake, Central Peru, vol.170, 2013, pp.433-451, Ioualalen M, Perfettini H, Condo S Y, Jimenez C, Tavera H, Figure 3)

Maule (2010)

On 27 February 2010, a great megathrust earthquake (M_w 8.8) occurred off the Maule region, central Chile, on the subduction zone interface between the converging Nazca and South American Plates. An ensuing tsunami struck the Chilean coast, extending ~ 300 km north and ~ 300 km south of the epicenter, taking almost 200 lives. More than 400 run-up measurements were collected with values ranging from 3 m up to ~ 30 m (Fritz et al. 2011b), distributed ~ 600 km along the coast between 33°S and 39°S . Many Chilean tide gauges recorded wave amplitudes between 1 and 2 m or more.

Significant tsunami waves were recorded around the Pacific Ocean, which is typical for megathrust earthquakes of this size. For example, the wave amplitude reached ~ 1.8 m at Hiva Oa Island of the Marquesas archipelago, French Polynesia, and ~ 1 m at several places in New Zealand, at Hokkaido in Japan, at the Kuril Islands, Russia, and at several locations in both California and Hawaii in the United States. The overall magnitude of the tsunami might have been reduced by the fact that a significant fraction of the coseismic displacement occurred inland (mostly subsidence) and not on the seafloor. Locally, subsidence enhances the tsunami impact.

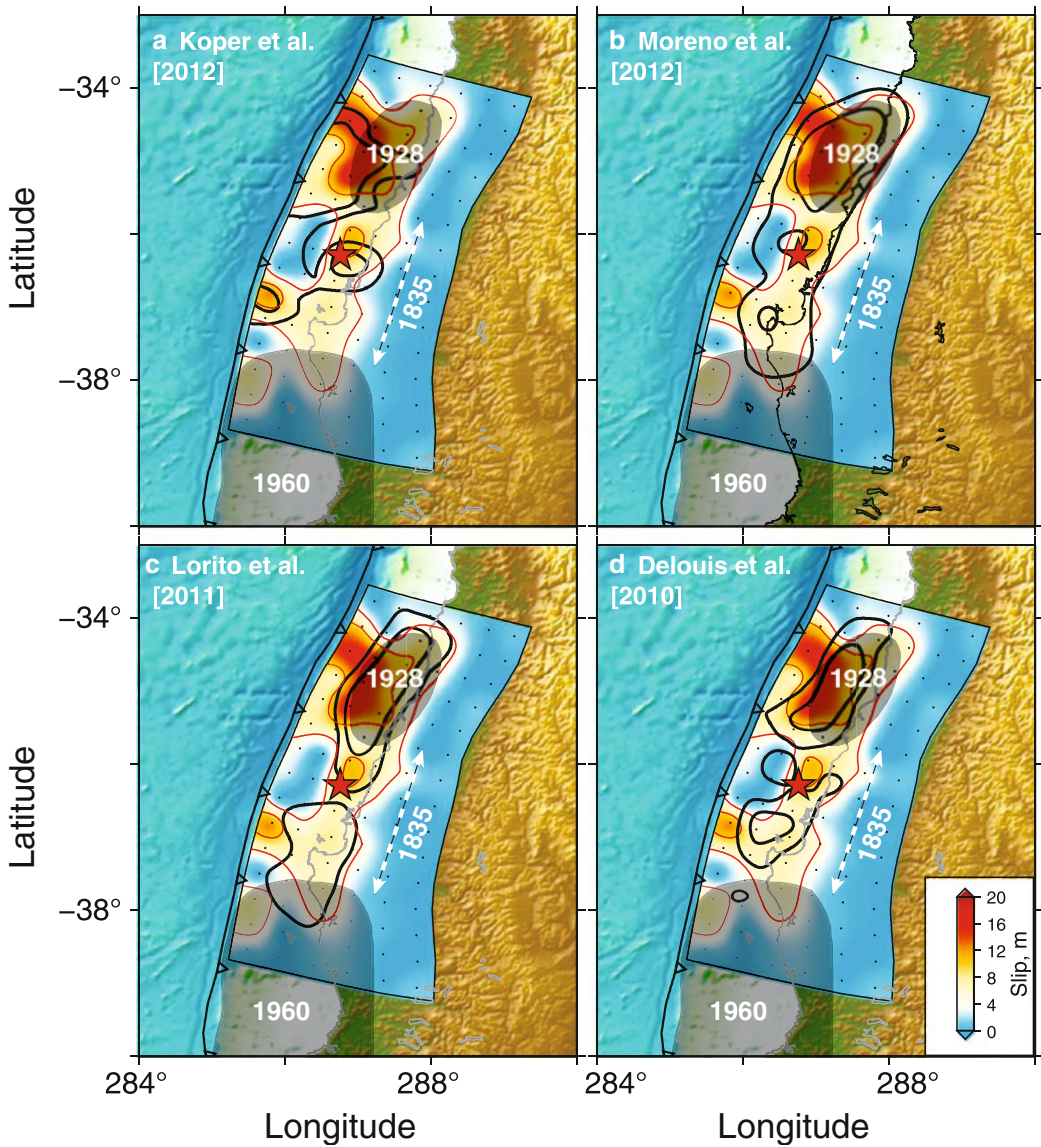
The seismic and tsunami source of the Maule earthquake has been investigated by a number of different groups (e.g., Delouis et al. 2010; Lay et al. 2010a; Lorito et al. 2011; Moreno et al. 2012; Koper et al. 2012; Lin et al. 2013), often using joint inversion of different geophysical datasets, including static data from GPS, InSAR, and land leveling, high-rate GPS, strong-motion, teleseismic body and surface wave data, and deep water and tide-gauge tsunami data.

To first order, all of the earthquake source models for the 2010 event feature a bilateral rupture extending north and south of the epicenter, with nonuniform coseismic slip along strike from ~ 38.5 to $\sim 34^\circ\text{S}$, a main patch of slip located offshore north of the epicenter with ~ 15 – 20 m of slip and a secondary patch to the south with lower slip ($< \sim 10$ m). Slip up-dip of the hypocenter, in the region between the two large-slip

patches, is small in most models. Most inversions do not have slip extending to the trench, although onshore geodetic data cannot constrain this well, and inversions of seismic data give variable placement of slip along depth.

Yue et al. (2014) performed a joint inversion of tsunami, geodetic, and seismic data, concluding that the rupture likely propagated up-dip to the trench along the two large-slip patches north and south of the epicenter (Fig. 9). Large outer-rise normal faulting aftershock seismicity is concentrated in the Nazca Plate seaward of these slip patches, as commonly observed seaward of large-slip regions in other shallow megathrust ruptures (2006 Java and 2010 Mentawai tsunami earthquakes; 2011 Tohoku and 2012 Haida Gwaii earthquakes). Resolution of the shallow megathrust slip was provided by the improved modeling of deep water tsunami signals, with propagation phase corrections accounting for two types of inaccuracy: neglect of seafloor elasticity and of density variations with depth in the ocean. These factors are particularly important for propagation to distant stations, as phase errors progressively accumulate, resulting in several minute too-early simulated arrival times with respect to observations. Without correction, these travel-time errors tend to shift inferred slip closer to the coast to delay the seaward arrivals. Other factors, such as imperfect bathymetry models, neglect of tsunami dispersion, and geopotential variations induced by the tsunami itself (e.g., Tsai et al. 2013; Watada et al. 2014) can also limit the precise determination of tsunami source area.

The 2010 Maule earthquake occurred in a region that had previously been identified as a seismic gap (Madariaga et al. 2010) between large ruptures in 1960 to the south and 1928 to the north. This portion of the plate boundary had last ruptured in an 1835 ($M \sim 8.5$) earthquake, an event that was experienced by Charles Darwin during his voyage on the *Beagle*. The “Darwin gap” (Lorito et al. 2011) was shown by GPS observations to be a region of slip deficit relative to plate convergence over the prior decade and inferred to have strong plate coupling (Moreno et al. 2010). The 2010 event nucleated within the



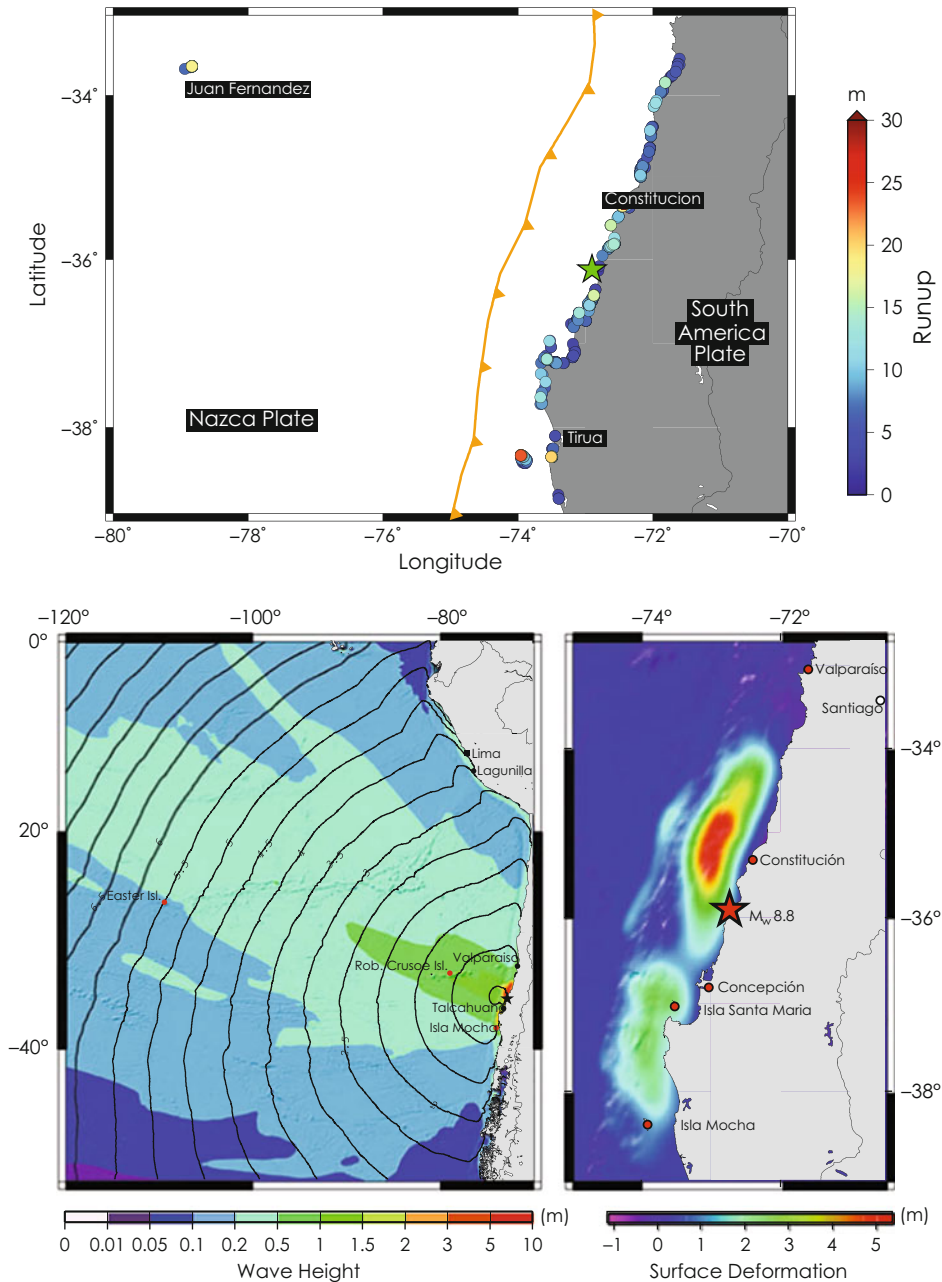
Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 9 Comparison of different slip distributions for the 2010 Maule earthquake

(Modified after Yue et al. 2014); dashed white arrow indicates the estimated 1835 earthquake rupture area, namely, the Darwin gap (Lorito et al. 2011)

gap, but the large-slip patches are not located within the presumed 1835 rupture area, with the northern patch overlapping the estimated rupture zone of the 1928 M_W 8.0 earthquake (Beck et al. 1998). Yue et al. (2014) note that many large aftershocks are located on the megathrust up-dip of the hypocenter, so there may be a mix

of stable and unstable friction in the central region of the gap.

As reported by Fritz et al. (2011b), the distribution of run-up measurements shows a broad range of values (3–30 m, Fig. 10), which are spatially well correlated with the along-strike-slip distribution. However, the maximum value



Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 10 Run-up distribution (*top panel*), maximum tsunami wave amplitudes over the South Pacific Ocean (*bottom-left panel*), and seafloor deformation field (*bottom-right panel*) for the 2010 Maule

earthquake (*bottom-left and bottom-right panels* from Fritz et al. 2011b) (Reprinted with kind permission from Springer Science+Business Media: Pure Appl Geophys, Field Survey of the 27 February 2010 Chile Tsunami, 168, 2011b, 1989–2010, Fritz H M, Petroff C M, Catalán, Cienfuegos R, Winckler P, Kalligeris N, Weiss R, Figure 10)

(~30 m) in Constitución, located on the coast near the main slip patch, involved a local effect of the wave surging up a coastal bluff. The run-ups decay both northward and southward but still show relatively high values (5–15 m) along the large-slip zone. On the other hand, the stretch of coast near the epicenter is characterized by relatively low run-up values, corresponding to the central region of low slip found in most published rupture models.

The inundation pattern did not involve large inundation distances, mainly due to the general steepness of the Chilean coast. The largest inundation distances were observed in Constitución (500 m) along the side of the city adjacent to the Maule River. Significant run-up values (~20 m) were observed along the coast of Mocha Island, located 30 km offshore, directly exposed to waves traveling from the southern large-slip patch. Similarly, large run-up values are recorded in Tirúa, a city located on the Chilean coast east of Mocha Island.

In the near field of the coseismic displacement, destructive late edge wave's arrival occurred some 3 h after the earthquake at Talcahuano harbor. This phenomenon is due to refraction and trapping of the energy along the continental shelf and slope, and both linear and nonlinear resonances of edge wave modes may amplify the tsunami. Edge wave excitation is a complex phenomenon related to fine slip distribution details (Geist 2009, 2012) and to coastal complexity (Yamazaki and Cheung 2011); thus, it is difficult to evaluate more precisely. Edge wave resonance likely also caused arrivals from multiple directions observed north of the epicenter as modeled by numerical simulations by Yamazaki and Cheung (2011). Significant edge wave excitation was observed also following the 2006 Kurils and 2009 Samoa earthquakes. One lesson from these events is that even if accurate modeling of tsunami edge waves may be difficult with conventional models (e.g., Liu et al. 1998), using too simplified earthquake slip and tsunami modeling may lead to severe underestimation for tsunami hazard assessments.

Finally, very high run-ups were reported along the coasts of islands comprising the Juan Fernández Archipelago, located more than

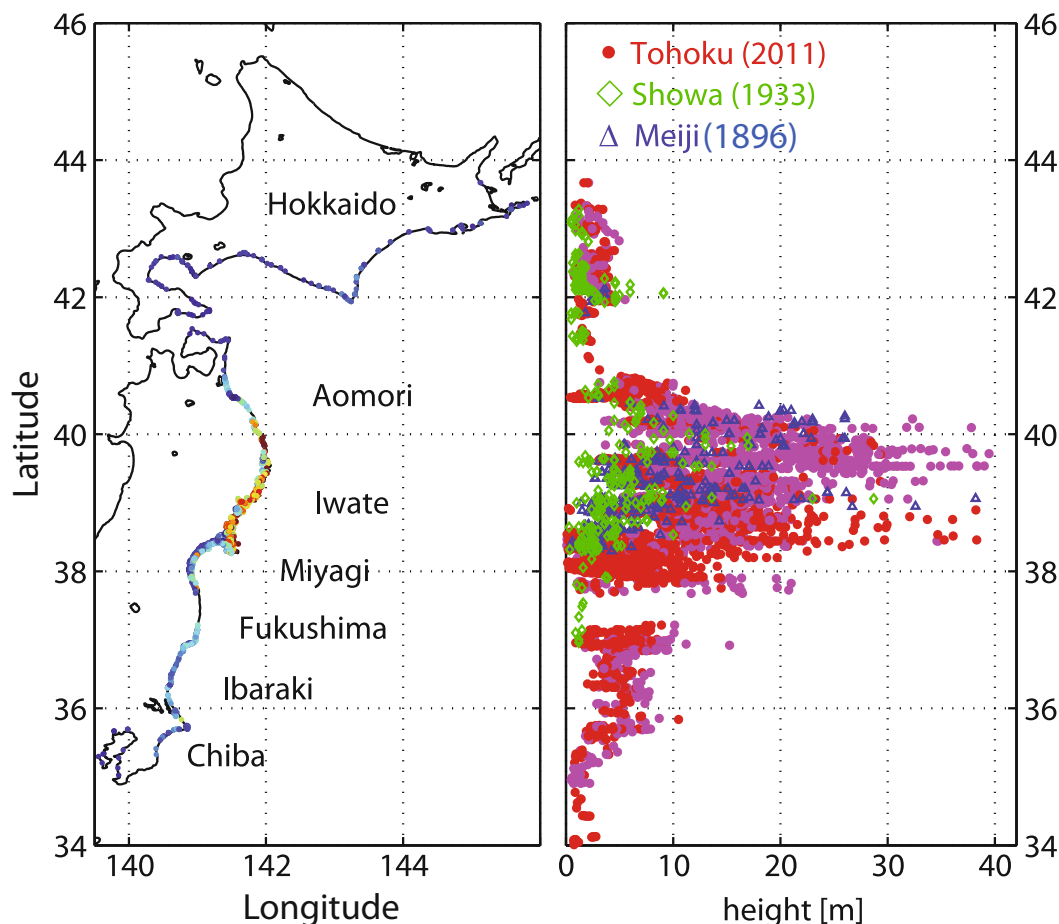
650 km off Valparaíso, in the perpendicular direction from the middle of the main slip patch. This direction is within the main lobe of the tsunami radiation pattern toward the Pacific Ocean, according to the simulation of Fritz et al. (2011b). Robinson Crusoe Island, impacted ~50 min after the earthquake origin time, first experienced a relatively slow-moving flood followed by a violent wave that destroyed most of the houses and local facilities. The maximum run-up measured was ~18 m, and the maximum inundation distance was ~300 m.

Tohoku (2011)

On 11 March 2011, a great earthquake (M_W 9.0) occurred offshore of the Tohoku region (Japan) on the subduction interface between the Okhotsk and underthrusting Pacific Plates. This is the largest recorded earthquake to strike Japan. It generated a huge tsunami that devastated the eastern coast of Honshu, killing at least 15,890 people and causing widespread destruction. It also precipitated a melt-down of the Fukushima Daiichi reactor, which, being classified as Level 7 on the IAEA event scale, is one of the two worst nuclear plant disasters ever.

Researchers from more than 60 universities and institutes participated in the post-tsunami survey that began only 2 days after the earthquake. They collected more than 5300 run-ups (Fig. 11) and flow depth observations along the eastern coast of Japan (Mori et al. 2011). A gap in measurements exists around the Fukushima area, which could not be surveyed due to the consequences of the nuclear accident. The run-up values range up to ~40 m along the coasts of Iwate prefecture (northwest from the epicenter). Sustained run-ups from 10 to 40 m occurred along a 300+ km stretch of coast; run-ups were generally ~2–3 m along the southern Honshu (Chiba) and Hokkaido coasts.

Comparable run-up heights along Iwate were observed in the 1896 Meiji and 1933 Sanriku-oki earthquakes (Fig. 11). Recorded earthquakes in the past few centuries off of central Honshu were less than magnitude ~8.2, much smaller than the 2011 event, and that had guided earthquake and tsunami hazard assessments, with the potential for



Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 11 Run-up distribution of the 2011 Tohoku earthquake (From Mori et al. 2011)

a much bigger rupture with very large tsunami along Sendai not receiving detailed consideration. Some recent work on the 869 earthquake had indicated run-up and inundation of the Sendai Plain that was to be matched by the 2011 event, but this had not yet penetrated into tsunami mitigation measures. Combined with other too simplistic assumptions, this reliance on recorded tsunami heights led, for example, to the deployment of 5.7 m tsunami walls near Fukushima (based on modern peak recorded tsunami along that stretch of coast that was caused by the distant 1960 Chile earthquake), which proved futile when the ~ 15 m wave generated by the 2011 event arrived (see discussion in Synolakis and K  no  lu 2015).

The extensive Japanese network of sea-level sensors recorded unprecedented tsunami observations. Many Japanese tide gauges measured tsunami amplitudes >10 m (up to ~ 18 m at Onagawa; Satake et al. 2013a); however, there is uncertainty regarding maximum values, as several tide-gauge facilities in the Iwate, Miyagi, and Fukushima prefectures were washed away or damaged during the tsunami inundation. Additionally, the tsunami was measured by several offshore GPS buoys, peaking up to 6 m along Iwate prefecture. Oceanward, deep water tsunami sensors recorded the first tsunami amplitudes >1 m. Wave amplitude exceeded 5 m at the sensor located ~ 100 km from the coast north of the epicenter and ~ 2 m at a DART buoy ~ 450 km

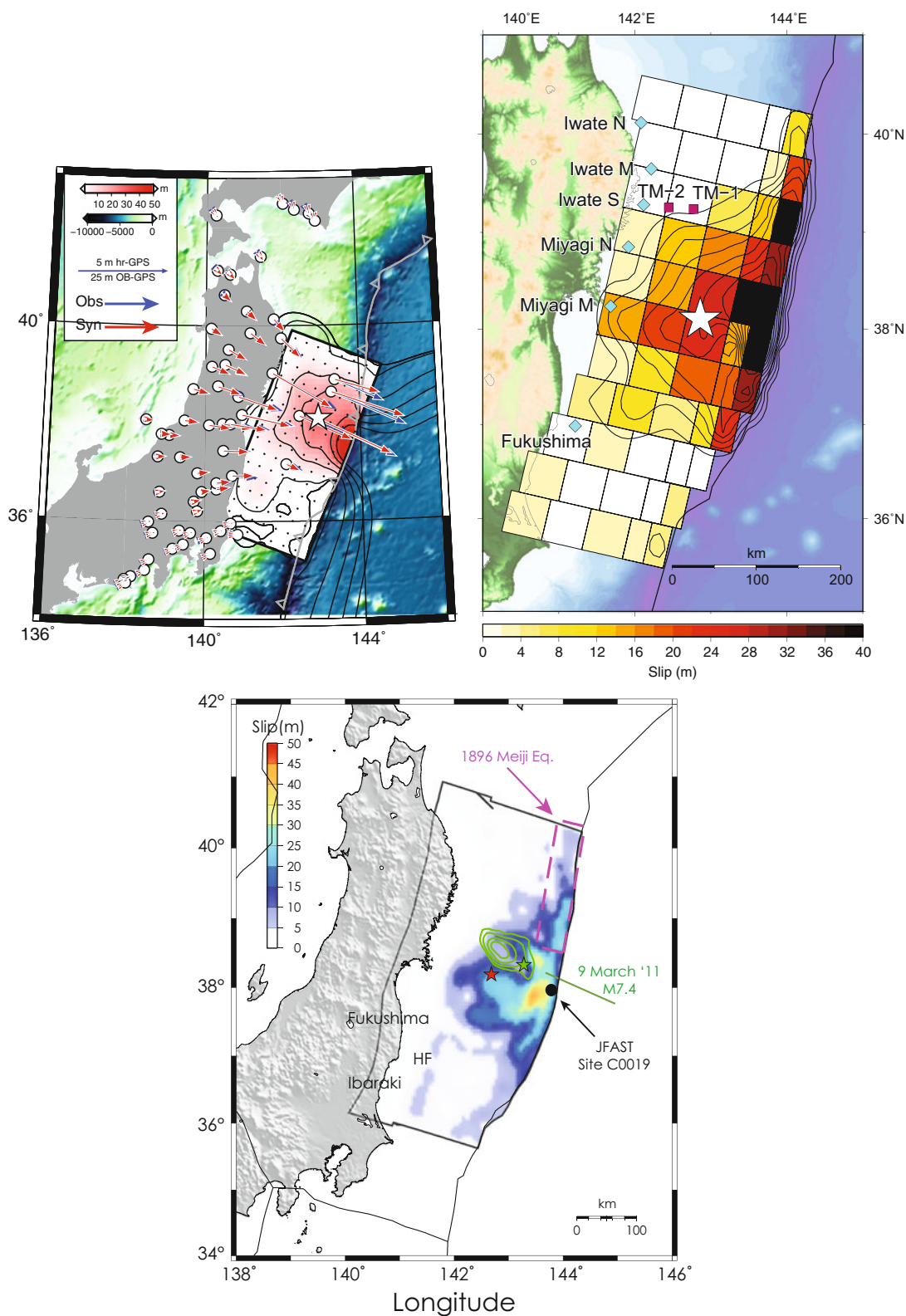


Fig. 12 (continued)

east of the Japan Trench. The tsunami propagated over the Pacific Ocean with wave amplitudes recorded by tide gauges up to ~ 1 m in Papua (Indonesia) and New Zealand, up to ~ 2 m in Hawaii, up to ~ 1.5 m at Nuku Hiva in French Polynesia, up to ~ 2.3 m in the Kuril Islands, and up to ~ 2.5 m in Chile and California (Crescent City).

The very dense Japanese geophysical networks also recorded an unprecedented amount of high-quality seismic and geodetic ground motion data in the near field of the source, and there is a superb global dataset of seismic observations, enabling many studies of this earthquake. Most of the more than 50 retrieved slip and rupture history models (e.g., Yokota et al. 2011; Lay et al. 2011b; Yue and Lay 2013; Satake et al. 2013a; Romano et al. 2014; Bletery et al. 2014, and many others, Fig. 12) share two first-order features: an overall spatially concentrated rupture, with a large-slip zone extending <200 km along strike at shallow depths with very large slip values (>50 m in numerous models) and extending <400 km along strike at greater depth with smaller slip. The entire 200 km wide expanse along dip of the megathrust appears to have ruptured, including regions of moderate size ruptures in the down-dip portion that failed in the last century. Models differ in whether and where significant slip extended to the trench.

Interpretation of the concentrated nature of the large-slip zone has been explored using seismic velocity anomalies around the fault plane. The slip in the central portion of the megathrust, where great earthquake ruptures typically occur (“Domain B” of Lay et al. 2012), corresponds well with a near-source volume characterized by high seismic wave velocity (Zhao et al. 2011) and high pre-event coupling (Loveless and Meade 2010). This region is confined between two low-velocity, low-coupling zones along strike; the latter may have acted as low strain or slip

velocity-strengthening barriers (Ye et al. 2012; Romano et al. 2014).

Extension of the 2011 rupture to the shallowest part of the megathrust, with peak slip (60–80 m) twice as large as that on the central megathrust, runs contrary to the common perception that the shallow subduction interface mainly experiences aseismic slip up-dip of the “seismic front,” a transition from seismic to aseismic behavior at a depth of ~ 10 km on the megathrust. However, the results of the JFAST drilling project which retrieved core from the Tohoku megathrust near the trench (Fulton et al. 2013 and references therein) suggest that, under conditions of high slip rate (e.g., Faulkner et al. 2011), the abundant smectite-rich clays on the megathrust may have exhibited slip velocity-weakening frictional behavior, with the huge slip possibly being also augmented by dynamic normal stress variation due to the reflection of seismic waves off the seafloor (Oglesby et al. 1998; Kozdon and Dunham 2013). Having large slip extend very close to the trench may have contributed to the huge intraplate extensional faulting sequence seaward of the trench within the Pacific Plate. Over one thousand normal faulting events occurred over a widespread area extending hundreds of kilometers from the trench (e.g., Lay et al. 2011a).

Most seismic and geodetic rupture models do not show significant shallow slip north of $\sim 39^\circ\text{N}$; however, the offshore tsunami buoy measurements off Iwate are compatible with little slip further north (e.g., Romano et al. 2012; Bletery et al. 2014). Accordingly, several models using tsunami data alone (Satake et al. 2013a) or in combination with static GPS offsets (Romano et al. 2014) in slip model inversions have inferred larger, delayed, or slow slip in the shallow sediment-rich megathrust (Satake et al. 2013a; Kubo and Kakehi 2013) extending to at least 39.5°N . This would overlap the estimated source area of the 1896 Meiji tsunami earthquake

Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 12 Slip distribution models of the 2011 Tohoku earthquake (*top-left*

panel, from Yue and Lay 2013; *top-right* panel, from Satake et al. 2013a; *bottom* panel, from Romano et al. 2014)

(Kanamori 1972), reinforcing the notion that the 2011 rupture was a combination of regular megathrust rupture and shallow tsunami earthquake. The northern extent of slip is important for evaluating whether the 1896 zone could produce another large tsunami earthquake. Some high-frequency-depleted seismic radiation from this region has been suggested (Yagi et al. 2012; Maercklin et al. 2012), although it is not evident in all back-projection studies, and any late moment release within the seismogenic passband should have been readily apparent in the rupture centroid time and long-period slip inversions.

Some authors (e.g., Tappin et al. 2014) argue instead that reconciling seismic and tsunami inversions with the large run-ups observed in the Sanriku region suggests the occurrence of a submarine landslide, whose scar would extend several tens of km along the strike direction, possibly corresponding to a feature mapped on the landward slope instead of near the trench axis (see also Satake and Fujii 2014). Quantification of the ~40 m run-up along the Sanriku Coast remains limited due to the lack of very high-resolution topo-bathymetric models, so the north-eastern extent of the 2011 rupture remains an open issue.

The large trench-reaching slip enhanced the huge tsunamigenic nature of this earthquake, with devastating effects and huge run-ups along most of the eastern coast of Honshu, broadside of the tsunami source. It is nevertheless interesting to decompose the slip distribution (Satake et al. 2013a) into its deep and shallow slip components (Fig. 13), in order to highlight a somehow counterintuitive feature, contradicting the commonly used simplification that the shallower the earthquake slip, the most damaging the tsunami will be. The relationship between source features and inundation particularly in the near field is more complex (e.g., Geist 2009). The Tohoku tsunami penetrated up to ~5 km inland along the Sendai and Ishinomaki plains. Simulating separately the tsunamis generated by the shallow and the deep slip shows that the latter generated an initial water displacement with larger wavelength (and thus longer tsunami periods) which can alone account for the observed inundation distance above the plains; on the other hand, the shallow

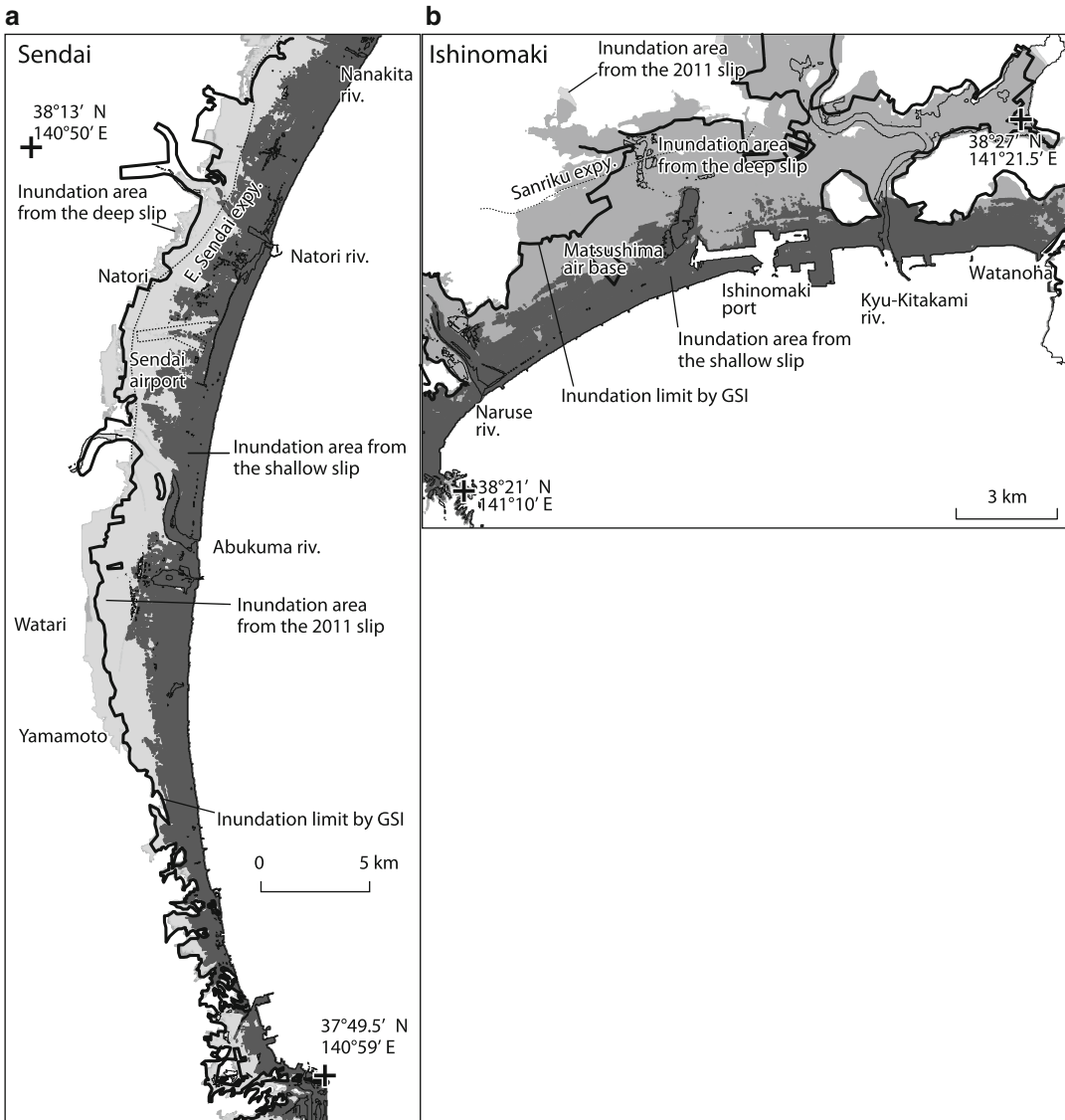
slip generates higher and steeper tsunami waves with more dissipative inland character achieving shorter inundation distances. This dual tsunamigenic slip behavior can be also observed in the tsunami waveforms recorded by the ocean bottom pressure sensors near the epicenter and at some GPS buoys located 20 km off the Iwate and Miyagi coasts, where both the shorter- and the longer-period components are clearly visible. Conversely, the largest run-ups have been observed in the near-oblique regime of the tsunami source, where the morphology of the Sanriku Coast contributed to the amplification of the tsunami waves.

Santa Cruz Islands (2013)

On 6 February 2013, a large underthrusting earthquake (M_w 8.0) occurred close to the Santa Cruz Islands (part of the Solomon Islands) on the plate boundary interface between the Australian and Pacific Plates. This is a particularly complex megathrust section, where plate geometry is not well known (Fig. 14). The hypocenter was located 76 km west of Lata, the main city of Nendo Island, at the northern end of the Vanuatu subduction zone where a transpressive transform fault extends westward to the Solomon Islands subduction zone. The 90° change in plate boundary strike produces a concentration of intraplate activity with normal faulting in the subducting Australian Plate and compressional and strike-slip faulting in the Pacific Plate. There had not been a prior major ($M > 7$) interplate earthquake in the vicinity of the 2013 event dating back to 1900.

This earthquake, the largest event to occur worldwide in 2013, generated a tsunami that struck the coasts of Nendo Island with maximum observed tsunami run-up > 11 m (Fritz et al. 2014) along the southwestern side of the island. The tsunami was recorded by tide gauges located in the Santa Cruz, Solomon, and Fiji Islands, with the maximum wave height (peak to trough) being recorded in Lata (> ~2 m); several NOAA DART buoys located in the southwestern Pacific Ocean spanning a wide range of azimuths recorded this tsunami too.

There are two published studies of the earthquake source, using teleseismic body wave



Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 13 Inundation

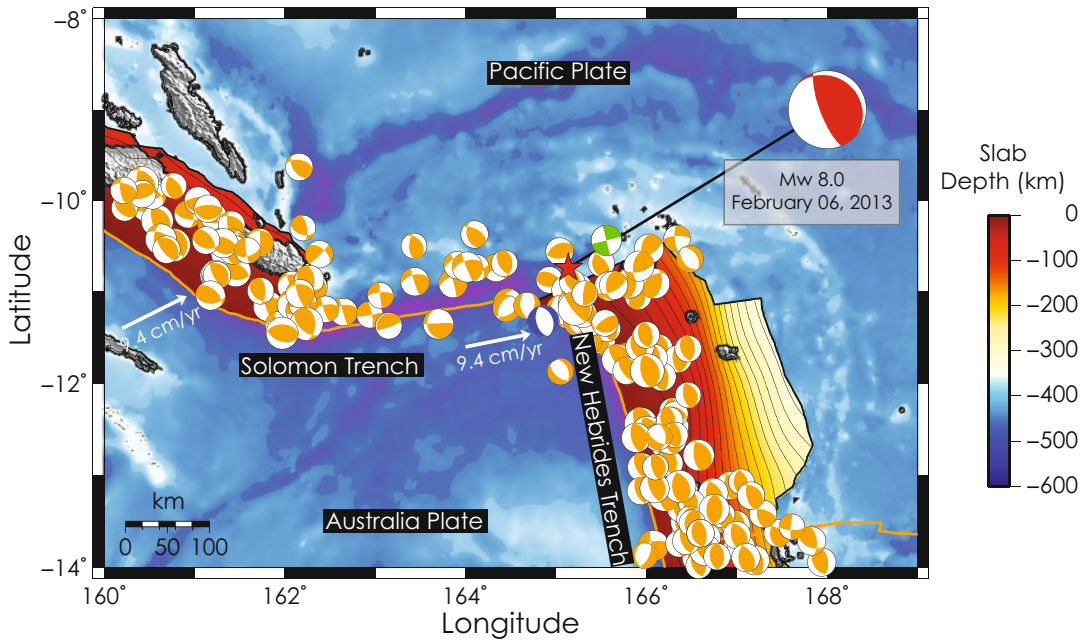
computed from deep (*light gray*) and shallow (*dark gray*) slip compared to inundation observed (*black line*) after the 2011 Tohoku earthquake (From Satake et al. 2013a)

inversion optimized by iterative forward modeling of deep-sea tsunami recordings (Lay et al. 2013a) and teleseismic body and surface waves (Hayes et al. 2014) and one of the tsunami source, from the inversion of DART and tide-gauge tsunami signals (Romano et al. 2015).

These models differ (Fig. 15) regarding slip amplitude, direction (rake), and distribution, as a result of using distinct datasets and fault model

geometries. This highlights the nonuniqueness of rupture models and their dependence on data choice and modeling assumptions.

The models of Hayes et al. (2014) and Romano et al. (2015) are essentially end members, as the first one is derived from ground shaking recorded at teleseismic distances, while the second one is derived by tsunami signals alone, although it imposes rupture velocity inferred from seismic



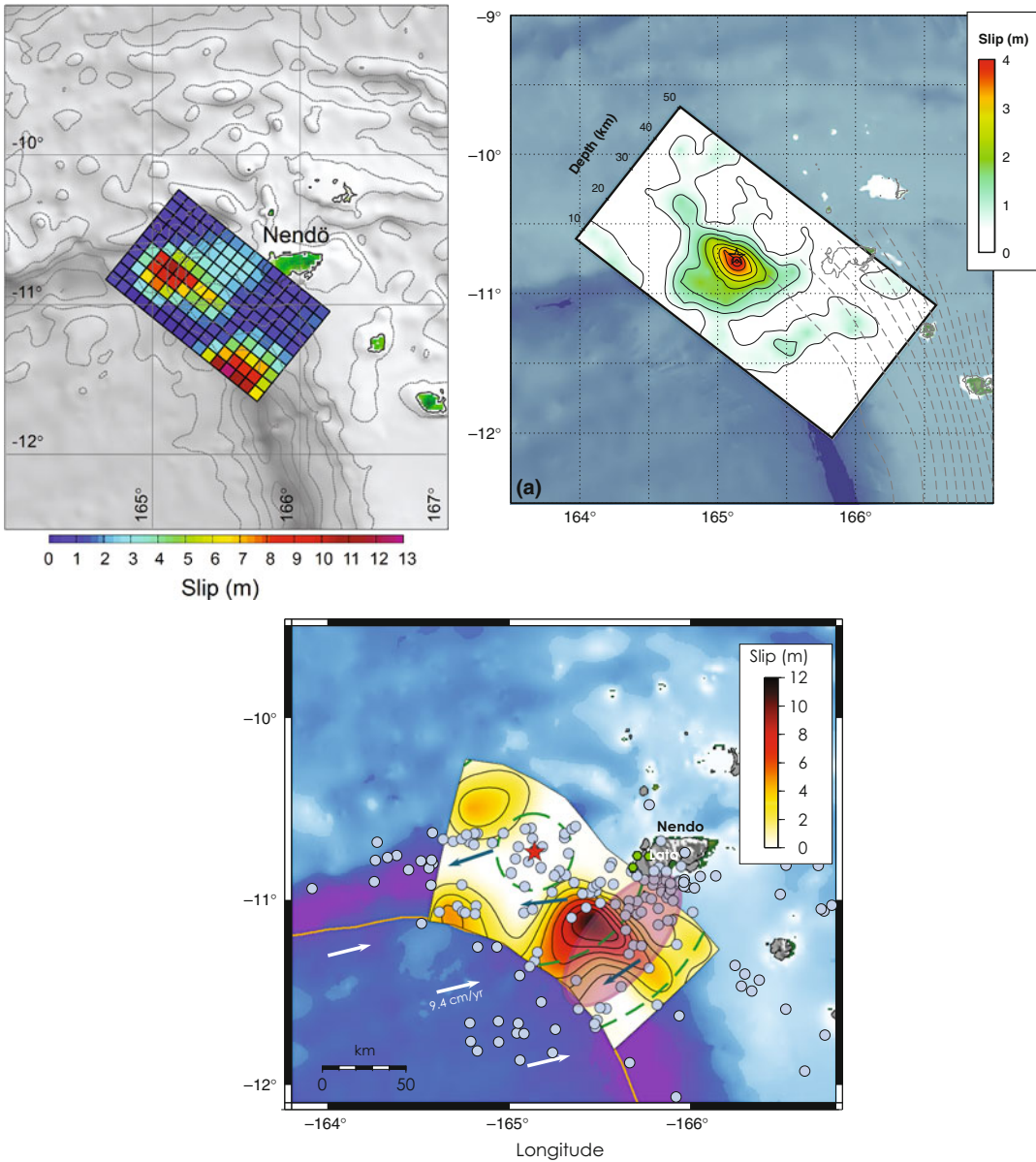
Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 14 Historical seismicity from 1976 and subduction zone features in the 2013 Santa Cruz Islands earthquake area, Romano et al.

2015 (With kind permission from Springer Science+Business Media: Nat Hazard Earth Syst Sci, Source of the 6 February 2013 Mw 8.0 Santa Cruz Islands tsunami, 15, 2015, 1371–1379, Romano F, Molinari I, Lorito S, Piatanesi A.)

modeling by Lay et al. (2013a). Intrinsically, these data types provide different sensitivity to the source process, and the models may not necessarily reproduce data that are not used in the inversion. The model of Hayes et al. (2014), characterized by moderate (<3.5 m) and quite deep slip, significantly underpredicts the observed tsunami amplitudes. Based on the analysis of aftershock locations and mechanisms, Hayes et al. (2014) speculated that there may have been some aseismic slip on the southeastern margin of their slip zone over hours or days; if this occurred fast enough ($<\sim 15$ min), it could have contributed to the tsunami generation modeled by Romano et al. (2015). The tsunami source model of Romano et al. (2015) will almost certainly not provide a good fit to the seismic data even though it assumed a reasonable rupture expansion velocity, as it places slip away from the hypocentral area where the models of seismic data place strong slip. However, it is possible to account for the tsunami generation and seismic waves with

a self-consistent model, as shown by Lay et al. (2013a).

The model of Lay et al. (2013a) should, in principle, offer the most complete image of the earthquake and its effects in terms of long distance shaking and tsunami generation, being determined using both kinds of data. Conversely, if aseismic tsunami generation occurred, it cannot be accounted for by forward modeling of the tsunami using the results of teleseismic inversion. A planar rectangular fault model is used, shifted shallower by ~ 10 km from that used by Hayes et al. (2014). Romano et al. (2015) use a 3D fault surface that may reduce uncertainty in tsunami modeling. However, the choice of model structure is very uncertain. Lay et al. (2013a) have reported rupture extending to the seafloor in the deepest part of the trench; the other models extend ~ 10 km further seaward, but without rapid shallowing of the plate boundary. It is not clear what choices are best.



Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 15 Slip distributions of the 2013 Santa Cruz earthquake (*top-left* panel, from Lay et al. 2013a (Reprinted from Tectonophysics, Vol. 608, Lay T, Ye L, Kanamori H, Yamazaki Y, Cheung K F, Ammon C J, The February 6, 2013 Mw 8.0 Santa Cruz Islands earthquake and tsunami, pp.1109-1121, Copyright 2013, with permission

from Elsevier); *top-right* panel, from Hayes et al. 2014 (Reprinted from Earth and Planetary Science Letters, Vol. 388, Hayes G P, Furlong K P, Benz H M, Herman H W, Triggered aseismic slip adjacent to the 6 February 2013 Mw8.0 Santa Cruz Islands megathrust earthquake, pp. 265-272, Copyright 2014, with permission from Elsevier); *bottom* panel, modified after Romano et al. 2015)

The low rupture velocity used in the models of Lay et al. (2013a), Hayes et al. (2014), and Romano et al. (2015) and the large slip (>10 m) at relatively shallow depth, extending up to the trench in some cases, are features that are typical of tsunami earthquakes. However, Lay et al. (2013a) find that the moment-scaled radiated energy is higher than typical of end-member tsunami earthquakes, suggesting that the event is a composite of typical interplate thrusting accompanied by some tsunami earthquake-like shallow slip.

The maximum observed run-up (~ 11 m) on the southwestern side of Nendo is ~ 1 km from the town of Malemgeulu and locates northeast of large coseismic slip (>10 m) regions in the models of Lay et al. (2013a) and Romano et al. (2015). This is likely a localized result of the direct arrival of the main tsunami waves generated by this patch with the coast lying broadside to the coseismic slip, particularly if the slip is dominantly in the thrust direction. Two additional relatively high run-ups were observed on the northwestern side of Nendo, 3 m along a stretch of coast facing the Pacific Ocean and 1.5 m near the city of Lata. The former could be due to the refraction of the tsunami waves traveling around the island from the southern slip patch, whereas the latter could be due to resonance effects within Graciosa Bay. An alternative hypothesis is that the main tsunami features are instead due to the hypocentral thrust slip patch in the model of Lay et al. (2013a), which they show is more tsunamigenic than the up-dip oblique-slip patch to the south.

Great Subduction Earthquake Doublets

When a large earthquake is soon followed by another nearby event with comparable size, the pair of seismic events is termed a doublet (e.g., Lay and Kanamori 1980). The second event in a doublet is essentially an unusually large aftershock, deviating from the common tendency for the largest aftershock to be about a magnitude unit smaller than the first event. Two great earthquake doublets occurred during the decade 2004–2013 involving unusual pairs of M_W 8 earthquakes

including strong outer-rise tsunamigenesis; the first one struck the central Kuril Islands region, with an M_W 8.3 plate boundary thrust earthquake on 15 November 2006 being followed by an M_W 8.1 outer-rise extensional earthquake on 13 January 2007; the second doublet was on 9 September 2009 at the northern end of the Tonga subduction zone and involved an initial M_W 8.1 extensional outer-rise earthquake followed within a minute by interplate thrust faulting with M_W 8.0. The recent Kurils and Samoa-Tonga events involve two of the larger recorded outer-rise normal faulting events. The 2004 (M_W 9.2) and 2005 (M_W 8.6) Sumatra great thrust earthquakes and the 2012 (M_W 8.7, 8.2) Indo-Australia intraplate strike-slip events are other recent examples of events with unusually large aftershocks.

Great megathrust events are commonly followed by normal faulting aftershocks in the outer rise (2011 Tohoku, 2010 Maule, and 2012 Haida Gwaii). The 2006–2007 Kuril doublet is the first documented case where the outer-rise-triggered event was also a great earthquake (the 1896 and 1933 Sanriku earthquakes could be another case, although the time separation of 37 years is longer than typically associated with doublets). The Samoa-Tonga doublet appears to be the first documented case for the opposite behavior, with megathrust faulting being dynamically triggered by an outer-rise “mainshock.”

Faulting interactions associated with doublets are important to consider for earthquake and tsunami hazard assessment after large events, although they tend to have low probability. There was high concern about the possibility of a 1933 Sanriku-like normal faulting event seaward of the 2011 Tohoku thrust zone and the associated danger of a second large tsunami impacting the damaged coast soon after the main rupture, but fortunately to date the extensional faulting occurred diffusely in many events on separate faults. Only a few previous documented cases of damaging tsunamis generated by outer-rise earthquakes exist; notably the 1933 Sanriku and the 1977 Sumba events (Kanamori 1971; Lynnes and Lay 1988).

Due to their relative infrequency, large outer-rise faulting is generally not considered in pre-calculated tsunami scenarios used for

operational forecasting, and such events represent a challenge for tsunami warning operations. Retrospectively, both the Kurils and Samoa-Tonga events would have been a challenge even for tsunami forecast based on fast moment tensors determination. In the first case, this is because of seismological ambiguity between the two conjugate steep focal planes, either southeast or northwest dipping, with strike direction ranging from subparallel to slightly oblique to the trench direction (Ammon et al. 2008; Lay et al. 2009). However, strain and accumulation release scenarios including earthquakes from 1963 to 2009 and the interseismic period seem to slightly favor a northwest dipping plane consistent with tensional stress induced by bending and downward pull of the subducting plate (Lay et al. 2009; Lay 2015). In the second case, the overall complexity of the sequence of earthquakes that occurred almost simultaneously and very close in space was hard to resolve (Beavan et al. 2010; Lay et al. 2010c; Satake 2010). Only very detailed analysis led to unraveling the complexity of the event that had initially been classified as a single earthquake with a complex moment tensor (Lay et al. 2010c).

Kurils (2006, 2007)

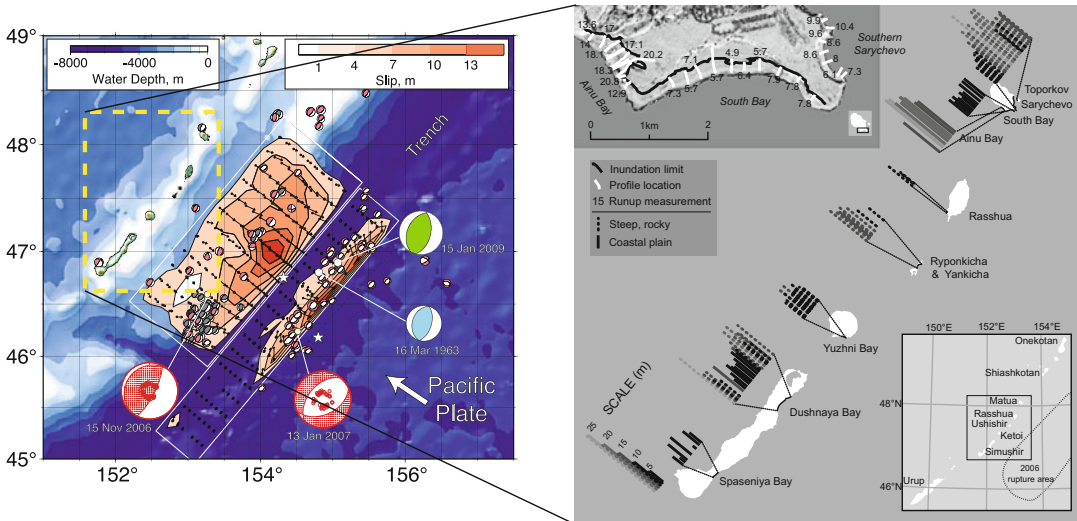
The great interplate M_W 8.3 Kuril Islands earthquake on 15 November 2006 occurred ~ 100 km southeast from the Simushir Island, of the Kuril archipelago. The hypocenter is located on the shallow part of the subduction interface between the Pacific and Okhotsk Plates. The earthquake nucleated between the rupture areas of the 1952 Kamchatka M_W 9.0 and the 1963 Kuril Islands M_W 8.5 earthquakes; in principle, this region could be considered a seismic gap. However, in contrast to some other cases (e.g., the Darwin gap in Chile), the lack of large recorded historical earthquakes and the presence of a large forearc basin unlike anywhere else along the arc (e.g., Lay 2015) made the seismic potential assessment of this area particularly uncertain. On 13 January 2007, a great M_W 8.1 normal faulting earthquake occurred in the outer-rise region of the 2006 event.

Both earthquakes generated tsunamis striking mostly uninhabited islands in the central area of the Kuril archipelago. The run-up measurements, collected during a post-tsunami survey conducted

a few months after the 2007 earthquake, highlighted an average run-up value of ~ 10 m estimated over a distance of ~ 200 km (MacInnes et al. 2009), with a maximum observed run-up of 22 m. Furthermore, the tsunami inundation distance varies from ~ 20 to ~ 500 m. Since the survey was conducted after the second of two events, it is difficult to clearly distinguish the causative source of the run-ups. It is likely that these observations are related to the first of the two events. MacInnes et al. (2009) argue that the observations are associated with the 2006 earthquake because at the time of the 2007 earthquake, Landsat images showed a snow accumulation in the Kuril Islands; under this condition, the tsunami inundation would not necessarily erode the coast making it difficult to clearly identify the run-up heights from debris deposits.

Both tsunamis propagated over the Pacific Ocean reaching Japan, New Zealand, French Polynesia, Hawaii, Chile, and the Western USA. Larger tsunami signals were recorded by the distant tide gauges for the 2006 earthquake than for the 2007 event, with the DART buoys documenting centimetric tsunami signals. For both earthquakes, the remote tsunami energy was mainly directed southeast (Rabinovich et al. 2008).

Interactions with submarine ridges and numerous seamounts located along the tsunami path were documented as sources of secondary tsunami arrivals larger than the direct wave for the 2006 tsunami (Kowalik et al. 2008). This can be clearly seen in the signals recorded by DART buoys D46404 and D46409, where late peaks appear, respectively, 2 and 3 h after the onset of the tsunami record. Accordingly, the initial wave at Kahului (Maui) of 50 cm amplitude was followed 2 h later by a wave of 76 cm amplitude, and similar patterns were observed elsewhere. An additional local amplification effect of note is along the Mendocino Escarpment, which entrapped, redirected, and amplified the tsunami wave heading to Crescent City (California). The small initial tsunami of ~ 20 cm was barely noticed there, whereas the highest wave of ~ 88 cm (~ 1.7 m peak to trough) was recorded 2–3 h later, resulting from a combination of distant and local amplifications. Similar anomalies at



Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 16 *Left panel*, slip distributions of the 2006, 2007 Kuril earthquakes (Modified after Ammon et al. 2008) (Adapted by permission from Macmillan Publishers Ltd: [Nature], Ammon C J, Kanamori H, Lay T, A great earthquake doublet and seismic stress transfer cycles in the Central Kuril Islands, Nature, 451:561–565, doi:10.1038/nature06521, Copyright 2008);

right panel, run-up distribution of the 2006 Kuril earthquake (From MacInnes et al. 2009) (Reprinted with kind permission from Springer Science+Business Media: Pure and Applied Geophysics, Field Survey and Geological Effects of the 15 November 2006 Kuril Tsunami in the Middle Kuril Islands, vol.166, 2009, pp.9–36, MacInnes B T, Pinegina T K, Bourgeois J, Razhigaeva N G, Kaistrenko V M, Kravchunovskaya E A, Figure 4)

Crescent City previously occurred for tsunamis from the great 1960 Chile and 1964 Alaska earthquakes.

The earthquake and tsunami source of these events have been investigated by inversion of teleseismic (e.g., Ammon et al. 2008; Lay et al. 2009) and tsunami (Fujii and Satake 2008; Baba et al. 2009) data. All models for the 2006 event show that most of the coseismic slip occurred in the shallowest part of megathrust with the rupture propagating northeastward from the hypocenter. The models inferred from teleseismic data have greater slip (>13 m, Fig. 16) than the model obtained by tsunami data inversion (~ 8 m). The rupture zone inferred by both models of the 2007 event is very narrow, and the slip distributions and northeastward rupture propagation are similar.

The run-up distribution associated with the 2006 event (Fig. 16) broadside to the tsunami source is fairly homogenous, with an average value of ~ 10 m; some particularly large values

(~ 20 m) have been observed on Matua and Simushir Islands, likely due to the coastal plain beaches that are backed by cliff or steep slopes. This morphological feature is common to most of Kuril Islands, and it is the cause of a limited inundation distance (~ 500 m); in addition, the presence of the islands is likely to have limited the tsunami penetration into and propagation within the Sea of Okhotsk.

Rabinovich et al. (2008) compared the observed and modeled tsunamis for both the 2006 and 2007 events. They demonstrated that the 2006 thrust earthquake resulted in larger edge waves than the 2007 outer-rise earthquake, as the former source was closer to the coast/shelf and aligned with it, consistently with the finding of Geist (2012) that this is a factor enhancing edge wave excitation. Edge waves can be also generated in the far field and in insular environments. They were detected after the 2006 earthquake also in Hawaii, where they were responsible for the highest surges (e.g., Munger and Cheung 2008).

Samoa (2009)

The great composite event of Samoa-Tonga occurred on 29 September 2009 in the southwestern Pacific Ocean near the Tonga Trench. The initial event epicenter was at a depth of ~ 20 km and located ~ 160 km southwest from the Samoa Islands. A normal faulting outer-rise event of M_W 8.1 occurred first, followed by two thrust sub-events on the nearby plate boundary megathrust located ~ 50 km south of the normal faulting.

The earthquake generated a tsunami that struck the nearby coasts of Samoa, American Samoa, and northern Tonga Islands. The run-ups along the coasts of Samoa vary from a few meters up to ~ 15 m, whereas they reach more than 20 m along the coasts of American Samoa and northern Tonga Islands (e.g., Okal et al. 2010). Run-ups up to ~ 10 m and ~ 15 m occurred along the southern side of the Sava'i and Upolu islands, respectively; the highest number of fatalities (~ 150 people) was reported in Upolu. Maximum inundation distance was 500 m along the Salani River in Upolu Island and 200 m in Taga (Sava'i Island). The tsunami had a significant impact on both the north and south side of Tutuila Island (American Samoa), with ~ 18 m in Poloa (on the western tip of the island) and 8 m in Pago Pago, the capital of the archipelago. The tsunami also propagated over the Pacific Ocean with decimetric wave amplitudes reaching the coasts of Chile, New Zealand, French Polynesia, Mexico, Peru, Hawaii, and California. In the near field, the largest tsunami waves were recorded by tide gauges located in Apia (Samoa, amplitude ~ 0.8 m) and Pago Pago (American Samoa, amplitude ~ 2.7 m).

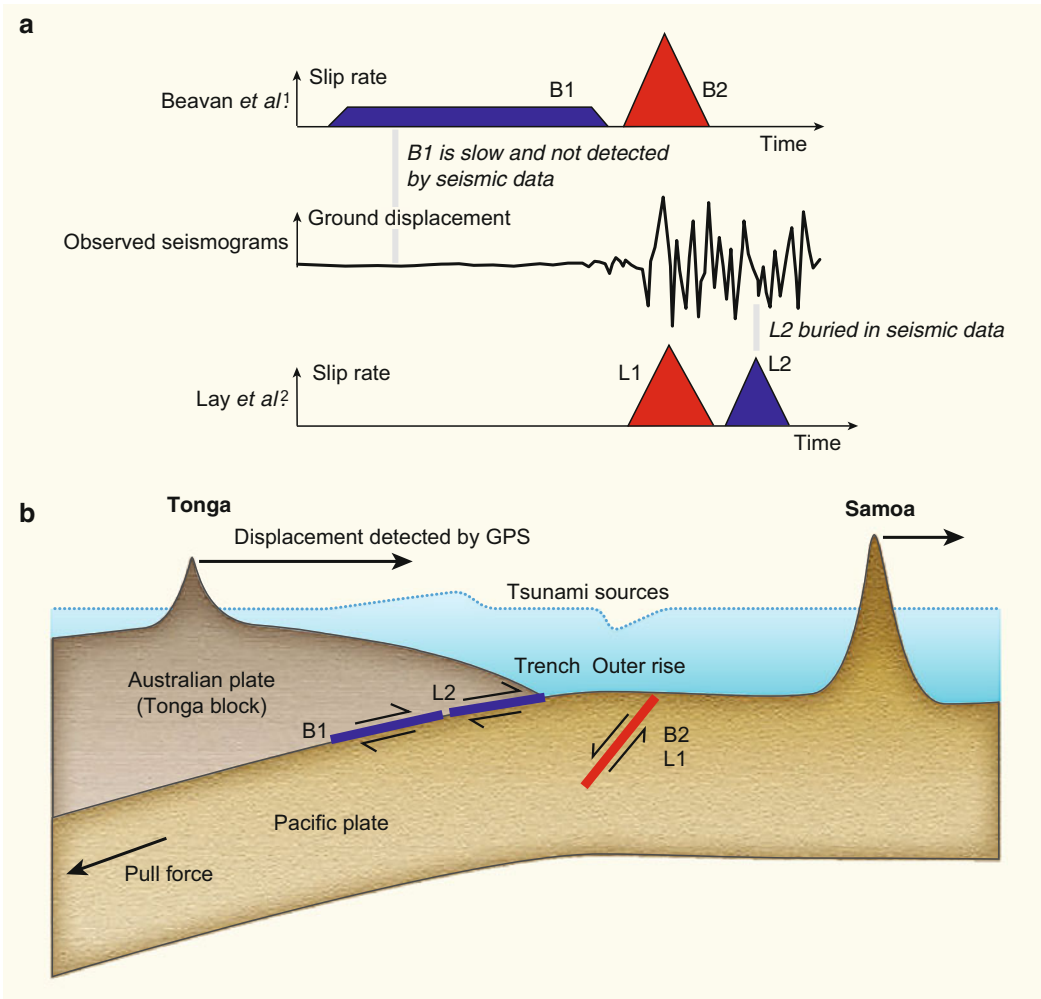
Beavan et al. (2010) (Fig. 17) embraced the hypothesis of an earthquakes doublet similar to the 2006–2007 Kuril, where the outer-rise event has been triggered in by an interplate rupture, based on the analysis of GPS and tsunami data. Eastward offset of a campaign GPS station on a northern Tonga Island is opposite to the expectation of only outer-rise normal faulting, and Beavan et al. (2010) postulated a thrust faulting event of about magnitude 8. Additionally, at some DART buoys, the signals are well reproduced only by considering a composite source, whereas

normal faulting is sufficient for reproducing the tide-gauge signal in Pago Pago in the near field (Fritz et al. 2011a). Since a thrust event was not immediately identified from the seismic signals, basing on tsunami and geodetic evidence, a slow thrust event was hypothesized. This interplate thrust earthquake could have then promoted normal faulting in the outer rise, thus making such a sequence of earthquakes similar to the 2006–2007 Kuril.

In-depth investigation of seismic waves recorded at teleseismic distances (Lay et al. 2010c, Fig. 17), supported by the back-projection analysis, demonstrated that thrust faulting actually followed the normal faulting. While there may have been some initial slow-slip thrust event that was not seismogenic, the moment of the triggered thrust faulting is sufficient to account for the GPS signal analyzed by Beavan et al. (2010), so there is no independent evidence for any slow-slip component to the process.

Samoa and American Samoa coasts were significantly struck by the tsunami (Okal et al. 2010; Fritz et al. 2011a, Fig. 18). In terms of source and propagation effects, tsunami simulations show that canyon-like bathymetric features offshore of Pago Pago could have driven the tsunami waves toward the city where the geometry of the harbor further amplified the tsunami, which penetrated 500 m inland (Okal et al. 2010).

The islands of the Tonga archipelago most impacted by the tsunami were Tafahi and Niuatoputapu. The largest run-ups (up to ~ 23 m) were measured on Tafahi, which is characterized by very steep coasts. The presence of coral reefs extending ~ 2 km offshore reduced the tsunami impact on Niuatoputapu, where the maximum run-up reached ~ 10 m; however, due to the flat coastal topography of Tafahi Island, there were larger inundation distances (~ 1 km). As for the Kuril 2006 event, edge waves have been documented and modeled, due to resonance with the shelf and embayment at Tutuila Islands, and they were likely responsible for the severest property damage and casualties that occurred during the 2009 tsunami (Roeber et al. 2010).



Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 17 Interpretations of the Samoa 2009 doublet

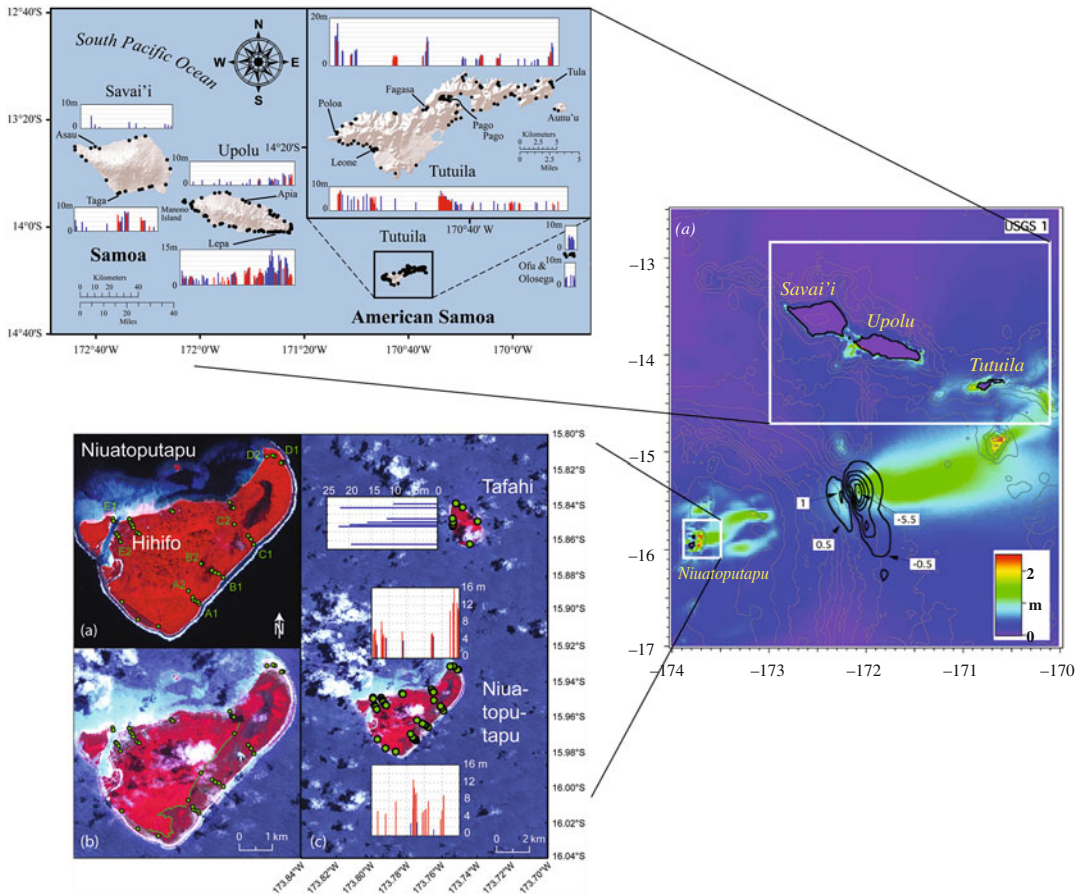
earthquakes (From Satake 2010) (Reprinted by permission from Macmillan Publishers Ltd: [Nature], Satake K, Double trouble at Tonga, Nature, 466:931–932, doi:10.1038/466931a, Copyright 2010)

Strike-Slip Events

Sumatra (2012)

On 11 April 2012, an M_W 8.7 earthquake occurred offshore of Sumatra, some 100–200 km southwest of the Sunda Trench, with epicenter ~ 300 km south from that of the 2004 Sumatra earthquake. It was followed 2 h later by an M_W 8.2 event with an epicenter ~ 200 km south. An M_W 8.7 intra-plate strike-slip earthquake had not been recorded before.

Due to its strike-slip character, involving limited vertical seafloor displacement, this earthquake generated a relatively small tsunami. The largest wave amplitudes were recorded by the tide gauges located in Indonesia (Meulaboh, ~ 1 m; Sabang ~ 0.4 m); the tsunami also reached the coasts of Thailand (Ko Taphao, ~ 0.1 m) and propagated over the Indian Ocean toward the Maldives Islands (Hanimaadhoo, ~ 0.3 m), Mauritius Islands (Rodrigues, ~ 0.2 m), the Seychelles Islands (Pointe la Rue, ~ 0.1 m), and La Reunion Island (Sainte Marie, ~ 0.2 m).



Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 18 Run-up distribution (left panels, from Fritz et al. 2011a (Reprinted from Earth-Science Reviews, Vol. 107, Fritz H M, Borrero J C, Synolakis C E, Okal E A, Weiss R, Titov V V, Jaffe

B E, Foteinis S, Lynett P J, Chan I, Liu P L F, Insights on the 2009 South Pacific tsunami in Samoa and Tonga from field surveys and numerical simulations, pp. 66-75, Copyright 2011, with permission from Elsevier); right panel, from Okal et al. 2010)

Earthquakes outside of subduction zones, or with unexpected mechanisms, are always a particular challenge for tsunami warning procedures. The Sumatra 2012 tsunami was initially overpredicted at several places, e.g., in Thailand and the Nicobar Islands, due to the assumption of a worst-case thrust event scenario and the early large magnitude estimates, before it was recognized that the event was a strike-slip earthquake. The warning was partially or totally adjusted by several warning centers as soon as focal mechanisms or tsunami measurements (e.g., at the DART buoys 23401 and STB01 midway between Sumatra and Sri Lanka) became available. False

alarms can always occur, and in a situation of large uncertainty immediately following a seismic event, erring on the cautious side may be considered better than an underprediction or a missing alarm. However, the rate of false alarms should be limited as much as possible since they undermine the credibility of warning systems which may limit the reaction of the population to the next warning, and avoiding unnecessary evacuations can save a lot of money.

It was initially speculated that the causative fault of the M_w 8.7 earthquake was aligned with the system of roughly NS trending crustal fossil transform faults off the Sunda Trench, within the

fossil fabric in the Wharton Basin, a diffuse zone of deformation at the boundary between the Indian and the Australian Plates, both converging under the Sunda Trench (Delescluse et al. 2012; Fig. 19).

Subsequent detailed source studies, mainly using teleseismic data, revealed a very complex multiple fault rupture process adding up to an M_W 8.7 event. It involved a complex system of orthogonal faults (Fig. 19), both with right-lateral WNW-ESE and left-lateral NNE-SSW strike-slip mechanisms (e.g., Yue et al. 2012; Wei et al. 2013; Hill et al. 2015). The maximum slip of ~ 37 m was observed on a fault centered around the hypocenter; the overall rupture zone, taking into account the entire fault system, spans ~ 300 km and extends in depth to 50–60 km. Using regional seismic data (Wei et al. 2013), a bilateral rupture along a NNE-SSW trending fault with maximum slip of ~ 8 m has been inferred for the M_W 8.2 aftershock.

The complex mechanism revealed by the source analyses is compatible with a reactivation/stress buildup process of this system of faults due to the elastic stress transfer following the 2004 Sumatra and 2005 Nias events and subsequent viscoelastic relaxation of the asthenospheric mantle occurred during the ~ 7 years in between the megathrust events and the intraplate sequence. The occurrence of strike-slip faulting rather than normal faulting as observed offshore of the Kuril Islands in 2007 and Samoa in 2009 is accounted for by the principle compressional stress lying in the Indo-Australian Plate due to the collision of India with Eurasia toward the northwest direction. Therefore, on a relatively short time scale, these events highlighted the interplay between megathrust events at the subduction interface and intraplate deformation within the incoming plates; on the longer time scale, the detaching process of Indian and Australian Plates is driven by their overall differential motion (Delescluse et al. 2012).

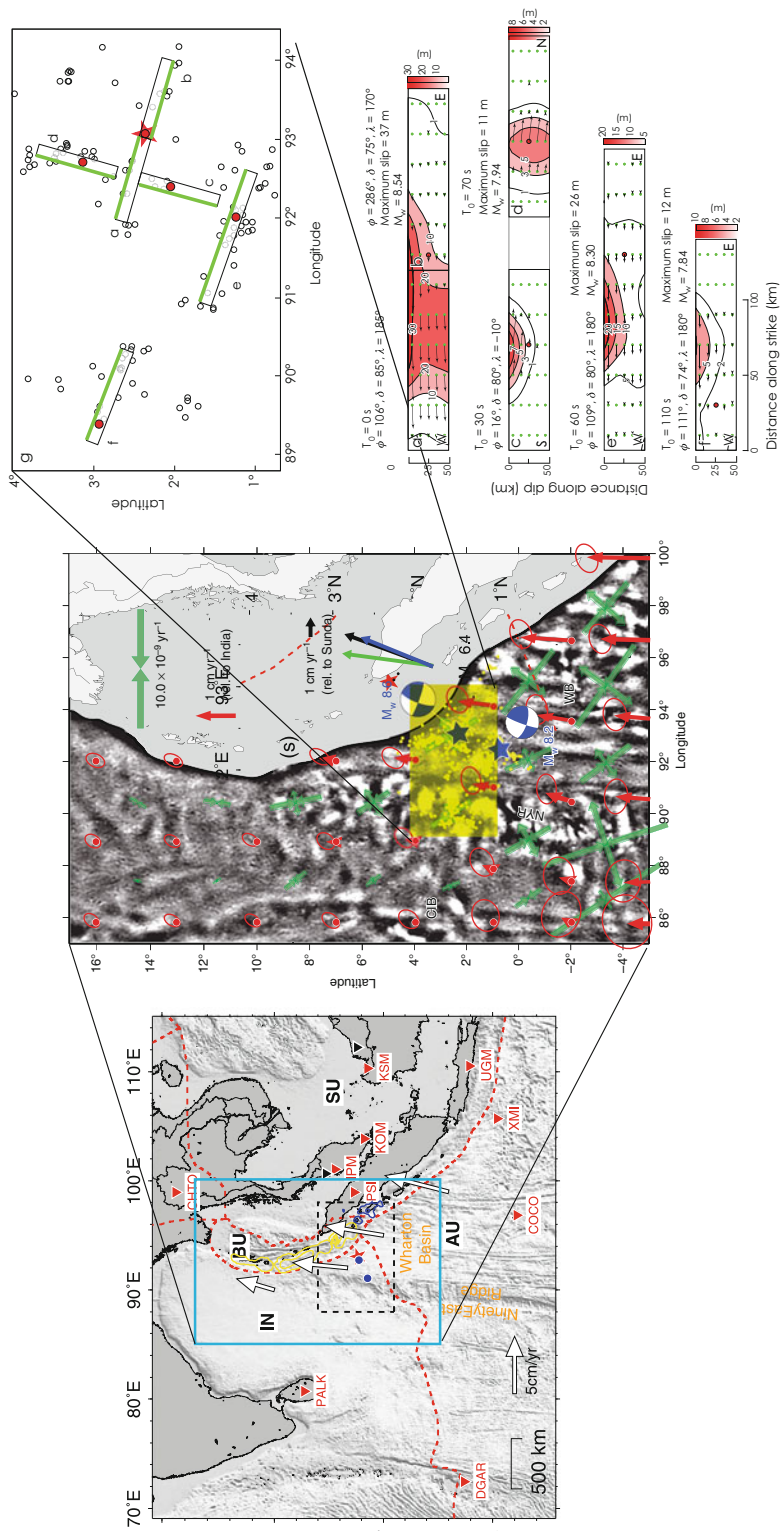
Tsunami Earthquakes

During the decade from 2004 to 2013, there were two major earthquakes along the Sunda Trench

that are classified as “tsunami earthquakes” and caused deadly and particularly damaging tsunamis along the coasts of Java and Sumatra. These are the 2006 Java and 2010 Mentawai earthquakes, both with M_W around 7.8. Prior events in this class have struck near Java in 1994 and near Sumatra in 1917.

Distinctive features of tsunami earthquakes are relatively long source durations, depleted short-period seismic source spectra, slow rupture velocities (~ 1.0 – 1.5 km/s), and deficiency of seismic moment-scaled radiated energy in comparison to other earthquakes (e.g., Polet and Kanamori 2009; this encyclopedia). Their mechanical properties are particularly intriguing as they do not involve large ruptures that begin deeper on the megathrust and then propagate to shallow depth like the 2011 Tohoku event. They occur entirely within the shallow and less rigid part of the megathrust, beneath the toe of sedimentary wedges. Many of the distinctive attributes of tsunami earthquakes can be qualitatively explained by the presence of low rigidity materials (e.g., wet, low seismic velocity sediments) that have greater slip for the same seismic moment than for high rigidity rocks like those deeper on the megathrust or in the surrounding crust (e.g., Geist and Bilek 2001). Large slip at shallow depth produces large displacement of the seafloor and efficient tsunami excitation near the deep trench (Geist 2009). Their occurrence in the vicinity of steep seafloor dip at the trench can also increase the net vertical motion of the seafloor resulting from horizontal displacement of the slope (Tanioka and Satake 1996).

The strong tsunami excitation for tsunami earthquakes results in greater coastal impact, particularly in the near field and at regional distances. Relatively weak ground shaking due to the depletion of short-period energy from the source can exacerbate the tsunami impact because people are not prompted to self-evacuate. It is of utmost importance to stress, during any tsunami preparedness campaign, that the duration of the shaking (even if it is weak) should be considered by coastal populations among the possible natural warnings of an imminent large tsunami. A sustained duration of about 1 min or more of shaking means: run to high ground!



Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 19 Tectonic framework and slip distribution of the 2012 great intraplate earthquakes; (left panel, from Wei et al. 2013; middle panel, modified after Delescluse et al. 2012 (Adapted by permission from Macmillan Publishers Ltd: [Nature], Yue H, Lay T, Koper K D, En Nature, 490:245–249, doi:10.1038/nature11492, Copyright 2012) boosted by the Banda-Aceh megathrust, Nature, 490:240–244, doi:10.1038/nature11520, Copyright 2012); right panels, modified after Yue et al. 2012 (Adapted by permission from Macmillan Publishers Ltd: [Nature], Yue H, Lay T, Koper K D, En Nature, 490:245–249, doi:10.1038/nature11492, Copyright 2012)

Java (2006)

On 17 July 2006 an M_W 7.8 earthquake occurred off the south coast of Java, at shallow depth on the subduction interface where the Australian Plate underthrusts the Sunda Plate approximately in the north-northeast direction across the Sunda Trench. This event generated a tsunami that killed more than 600 people in Java.

Very weak ground shaking preceded the large tsunami arrival. Permisan experienced 21 m peak run-up along a 1–2 km long stretch of coast, with an inundation distance of ~ 1 km. This was much larger than the average run-up value of 5–7 m along ~ 250 –300 km of the southern Java coast (Fritz et al. 2007). This tsunami was clearly recorded at tide gauges around the southeastern Indian Ocean. Some inundation occurred, close to high tide, on the Western Australian coast, peaking at ~ 7 m on a steep cliff within Shelter Bay. However, tsunami effects were comparatively larger in the near field than in the far field with respect to those of the tsunamis generated by larger magnitude earthquakes.

The slip distribution of the 2006 event has been studied using body and surface wave (Ammon et al., 2006, Fig. 20) and tsunami (Fujii and Satake 2006) data, respectively. Both studies infer a 200–250 km rupture extent expanding unilaterally southeastward from the epicenter, with slip mostly confined to the shallowest part of the megathrust, up-dip of the 15 km hypocenter. The low estimated rupture velocity (1–1.5 km/s), very long duration (~ 190 s), and compound spectral features of the source radiation, with high frequencies originating from localized regions, and long-period waves excited by the entire rupture, along with the large shallow megathrust slip and large tsunami excitation, identify the 2006 Java event as a tsunami earthquake.

Significant differences exist in the inferred slip on the fault; the teleseismic inversion estimates two to three times larger localized slip values relative to the coarser model used in the tsunami inversion. This appears to be largely due to the low, 10 GPa rigidity value, typical of sedimentary materials, used to convert seismic moment to slip for the teleseismic inversion. The actual rigidity is not well known, and large

variations are common in shallow subduction zones. On the other hand, the tsunami recordings used for inversion are mostly obtained from tide gauges located at regional and far-field distances, and consequently the source slip is “averaged” over 50×50 km² patches, because only near-field tsunami data are sensitive to finer scale slip heterogeneity.

The length of the Java coast hit by the tsunami inundation is consistent with the along-strike extent of the inferred source, to which it lies broadside. The tsunami earthquake nature of the Java event, with large shallow slip, accounts for the damaging tsunami, whose consequences, mercifully, were likely limited by the fact that it occurred on Monday afternoon, when there were very few people on the beaches.

The run-up “outlier” observed at Permisan led to the suggestion of a possible additional non-seismic source for this tsunami, namely, a submarine landslide (Fritz et al. 2007; Fig. 20). However, a combination of the source features and the tsunami propagation toward land offers a convincing alternative explanation. The convex shape of the subduction as seen from the Java coast may have led to a strong focusing effect, without ruling out additional focusing and amplification effects due to the diffractive behavior of the bathymetric features which are not known in detail (Kânoğlu et al. 2013 and references therein).

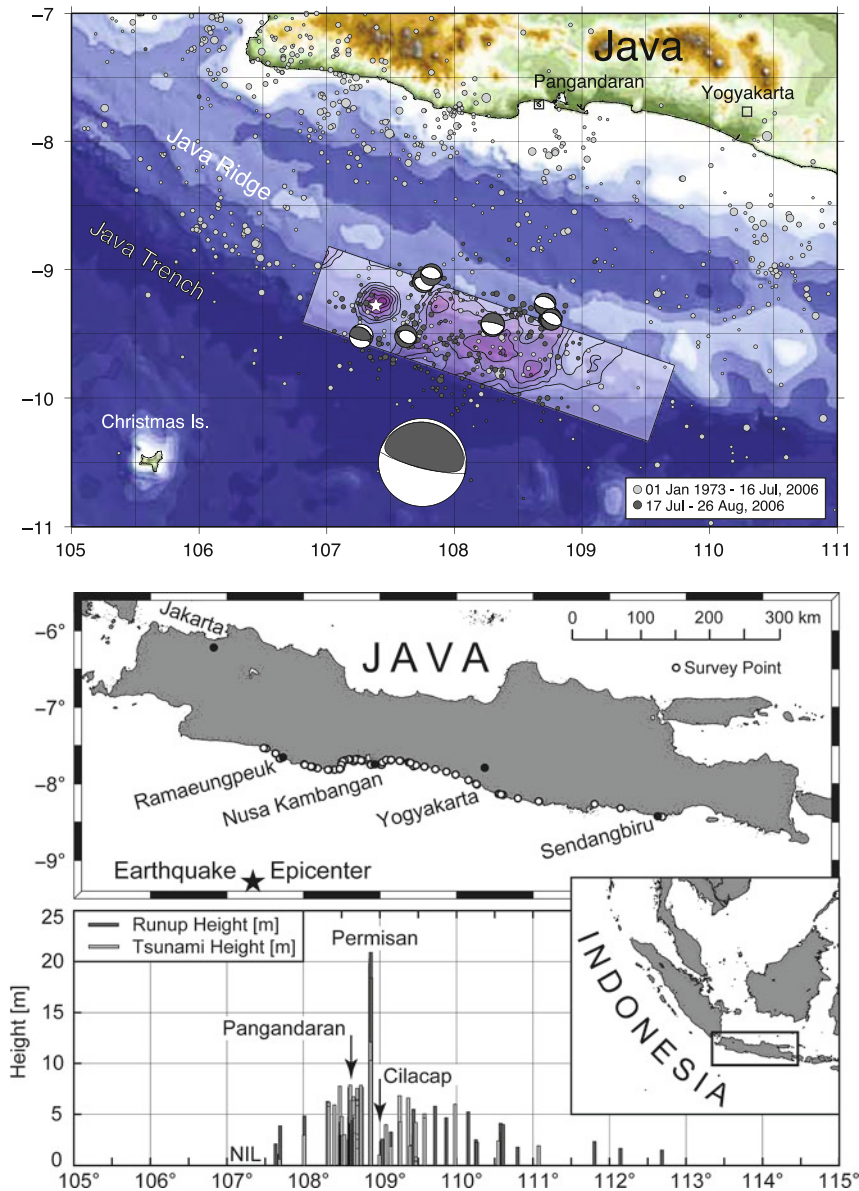
Mentawai (2010)

About 4 years after and ~ 1000 km northwest along the Sunda Trench from the Java 2006 event, another large (M_W 7.8) underthrusting tsunami earthquake occurred on 25 October 2010 offshore from the southeastern Mentawai Islands. Including the 1994 Java tsunami earthquake which occurred ~ 600 km to the east of the 2006 Java event, this was the third tsunami earthquake in 16 years along the same megathrust. The hypocenter of the 2010 Mentawai event was located ~ 40 km offshore southwest from the South Pagai Island, just up-dip of the rupture zone of the 12 September 2007 Bengkulu M_W 8.4.

The 2010 earthquake generated a large tsunami that struck the west coasts of the Mentawai

Islands, with the strongest effects on North and South Pagai Islands (Hill et al. 2012). In particular, the run-up observations vary over a range of 2–17 m with peak values along the mid-west coast of South Pagai Island. The death toll following this tsunami exceeded 500. A tsunami warning was issued that did not reach the majority of the

population due to the absence of sufficient infrastructure for alert dissemination. Again, self-evacuation was limited: the earthquake produced weak ground shaking and was perceived as being much weaker than the 2007 Bengkulu earthquake, which generated a relatively modest tsunami, or the 2009 intermediate depth Padang earthquake,



Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 20 Slip (top

panel, from Ammon et al. 2006) and run-up distribution (bottom panel, from Fritz et al. 2007) of the 2006 Java earthquake

which only generated small anomalies recorded by the Padang tide gauge.

The tsunami has been measured offshore by buoys (DART and a nearby GPS buoy which recorded the tsunami from 10 min after the earthquake) and tide gauges spread around the Indian Ocean. Most recorded relatively weak, but distinct, centimetric and decimetric tsunami signals that have been used by different groups for inversions or for calibration of source models were obtained by teleseismic inversions.

Several source models have been proposed for the 2010 Mentawai earthquake. Even though retrieved using different kinds of geophysical data, they all share a common pattern of slip mostly confined along the shallowest part of the megathrust up-dip of the epicenter (Newman et al. 2011; Lay et al. 2011c; Hill et al. 2012; Satake et al. 2013b; Yue et al. 2015). The most recent and best constrained model (Yue et al. 2015), obtained by joint inversion of tsunami, regional high-rate GPS, and teleseismic data, indicates that the rupture propagated up-dip and northward from the epicenter and that most or almost all the slip was released in a very narrow patch close to the trench, elongated along the strike direction, with an almost purely thrust slip direction.

The large (up to 15 m, Fig. 21) shallow slip can account for the run-up observations on the

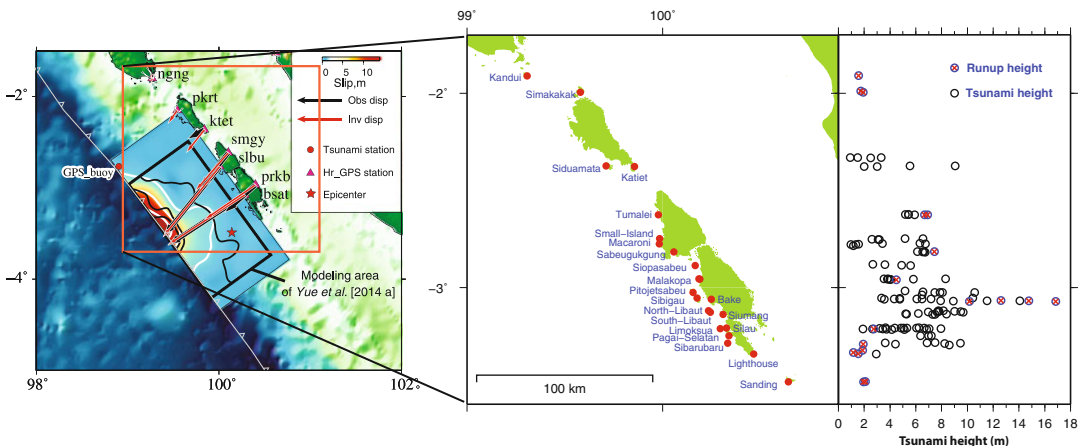
Pagai Islands (Hill et al. 2012; Satake et al. 2013b). For example, the tsunami penetrated up to ~ 400 m inland on Muntei (North Pagai Island) and up to ~ 300 m on Maonai (South Pagai Island), likely due to the U-shaped morphology of these bays. In addition, the maximum run-up measurement (~ 17 m) for this event was observed on Sibigau, located ~ 5 km off South Pagai and closer to the tsunami source; on this island, the tsunami penetrated up to 600 m inland. Conversely, the Pagai Islands acted as a natural shield greatly mitigating the tsunami impact on cities like Padang on the Sumatran coast to the northeast.

The limited impact of the Java and Mentawai tsunami earthquakes in the far field, with respect to other events induced by earthquakes of larger magnitudes, demonstrates once more that local tsunamis are greatly influenced by the features of the slip distribution, while distant tsunamis are controlled to first order by the earthquake magnitude (e.g., Okal 1988; Geist 2009).

Special Cases

Solomon Islands (2007)

The Solomon Islands-New Britain subduction zones have hosted numerous large earthquakes,



Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 21 Slip (left

panel, from Yue et al. 2015) and run-up distribution (right panel, from Hill et al. 2012) of the 2010 Mentawai earthquake

often occurring in triggered doublets or multiplets (e.g., Lay and Kanamori 1980), and some events have been tsunamigenic (e.g., Geist and Parsons 2005) with impact on the adjacent islands. The central Solomon Islands subduction zone is the site of a triple junction, with the Australian and Solomon Sea Plates underthrusting the Pacific Plate. There is slight divergence of the two subducting plates, and they are separated by a ridge-transform system with very young (<3.5 Ma) lithosphere entering the trench. Historical seismicity near the triple junction has been low, and there is a small slabless window at depth, so even intermediate depth seismicity is low in the vicinity of the triple junction. Conventional ideas about subduction of very young, thin lithosphere had suggested that seismic coupling might be low in the region, accounting for the lack of any large historic earthquakes, but this was proved incorrect when a major rupture stretched right across the triple junction in 2007 (Furlong et al. 2009).

On 1 April 2007, a great interplate earthquake occurred off the New Georgia Group of islands (Solomon Archipelago) with magnitude M_W 8.1. The hypocenter is shallow on the subduction interface between the Australian and Pacific Plates, close to the triple junction (Fig. 22), but the rupture extended northwest, crossing over the triple junction, with large slip also occurring on the interface between the Solomon Sea and Pacific Plates. This surprised many researchers.

This earthquake generated a tsunami that struck most of the islands in Western and Choiseul provinces of the Solomon Archipelago. In general, when a typical megathrust earthquake occurs very close (a few tens of km) to the coast, then the “lead time” between detection of the shaking and the tsunami impact can be very short. Luckily, in this case, the message “run to high ground after an earthquake” had been passed on to younger generations by the survivors of prior tsunamis, and this triggered an effective self-evacuation that, despite massive tsunami impact (more than 6000 damaged or destroyed houses), limited the death toll to less than 60 fatalities (Fritz and Kalligeris 2008).

Several source models of the Solomon Islands earthquake have been published. For example,

based on inversion of teleseismic body waves (Furlong et al. 2009, Fig. 22) or coral uplift and coastal subsidence data (Chen et al. 2009). Furlong et al. (2009) inverted global seismic body waves, finding a unilateral rupture that propagated northwest from the epicenter at a relatively high rupture velocity of 2.5 km/s. Their model features two shallow slip patches, straddling the junction. The rake angle ($\sim 90^\circ$) of the northernmost patch (maximum slip ~ 7 m) is consistent with the convergence direction north of the junction, whereas the southernmost patch (maximum slip ~ 4 m) is consistent with the oblique convergence direction (rake $\sim 50^\circ$) of the Australian Plate south of the junction. This suggests that the event involved two plates underthrusting a third plate in different directions during a single earthquake, consistent with the tectonic setting.

The pattern of slip found by the analysis of geodetic data is to first order consistent with the teleseismic model but features larger slip values (up to 30 m locally) over a narrower zone at very shallow depth; this difference might be due to postseismic uplift being included in the static offset data used, as the campaign GPS sites were reoccupied 1 or more weeks after the earthquake. Also, the geodetic data have diminished resolution of slip as distance from the coast increases. There is also uncertainty in the value of rigidity used to convert seismic moment into slip in seismic inversions. Nevertheless, this is the first well-documented case of a great event rupturing through a triple junction. Together with other cases, such as the 2002 M_W 7.9 Denali earthquake, in which a complex rupture involved three different faults and both a thrust and a strike-slip mechanism (Eberhart-Phillips et al. 2003), this earthquake is a further caveat against assumptions of strong segmentation based on the presence of structural barriers in hazard analyses (Taylor et al. 2008), as discussed for several events analyzed in this entry.

The post-tsunami surveys also measured tsunami run-up (Fig. 22). The largest impact occurred on the Simbo, Ghizo, and Ranongga Islands which are located within the vertical coseismic displacement field; the maximum run-up (~ 12 m) was measured on Simbo Island

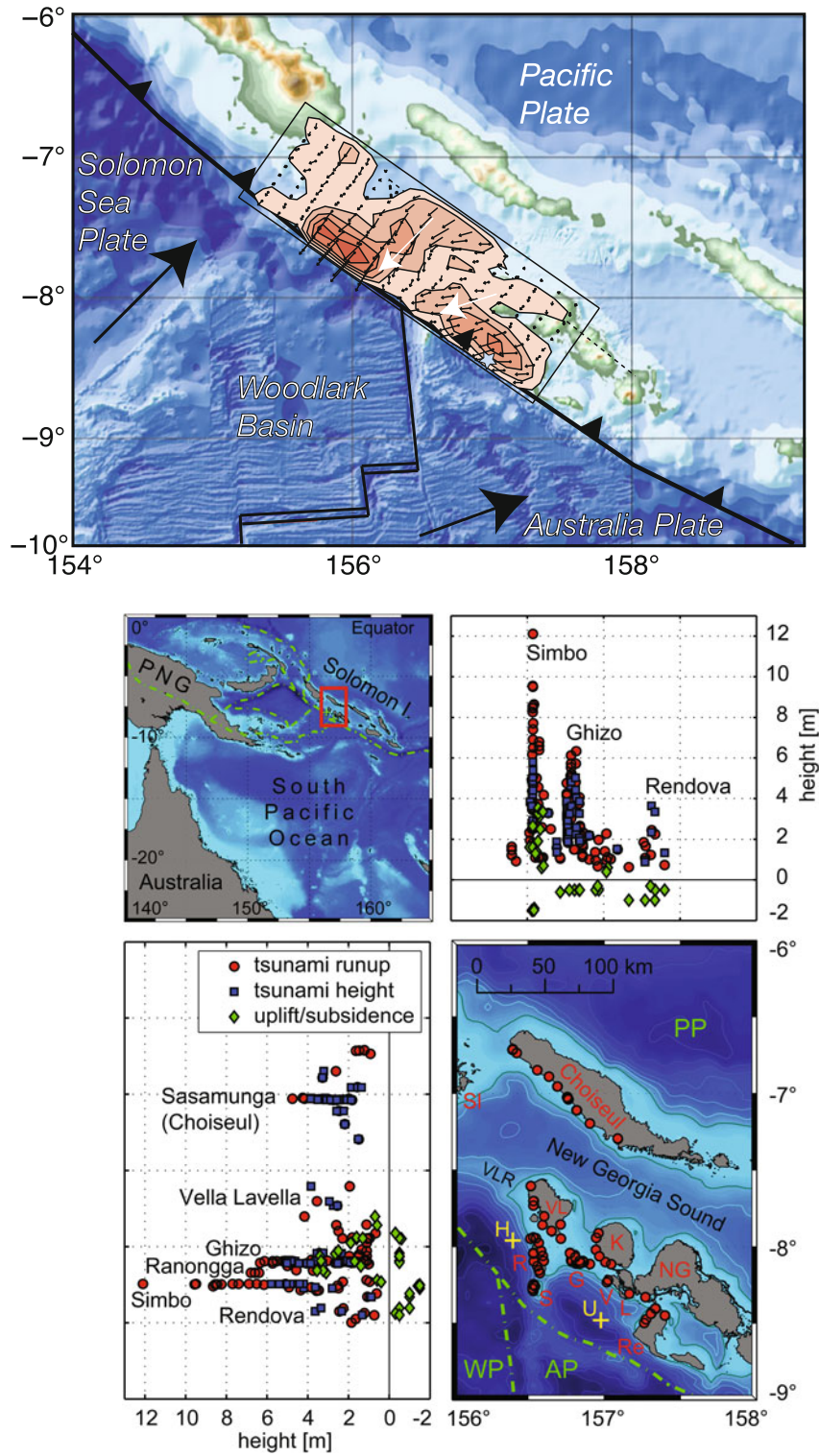


Fig. 22 (continued)

and is consistent with the shallow slip near the trench (Chen et al. 2009), whereas ~ 7 m have been measured on Ranongga Island and ~ 6 m on Ghizo Island. Run-ups of ~ 5 m are reported along the coasts of Choiseul Island located in front of the northernmost and largest patch of slip. The presence of coral reefs and the straits that separate the New Georgia islands is likely to have attenuated the tsunami impact on these islands (Fritz and Kalligeris 2008).

Haida Gwaii (2012)

On 28 October 2012, a major (M_W 7.8) shallow thrust earthquake occurred along a transpressive boundary of predominantly right-lateral strike-slip motion between the Pacific and North American Plates. The thrusting represents slip partitioning in a restraining bend of the Queen Charlotte Fault, which has pure strike-slip motion further to the NNW. There is uncertainty in the precise amount of convergence over the past 6 Ma, during which the relative plate motion between the Pacific and North American Plates has changed slightly, but it is likely that the Pacific Plate underthrusts to at least the eastern side of the southern island of Haida Gwaii (Moresby Island) (Fig. 23).

The earthquake hypocenter was located close to the Pacific coast beneath western Moresby Island, in the Haida Gwaii archipelago, Canada. The accompanying tsunami struck primarily on the nearby western coast of the island, which shielded the coast of British Columbia. During a post-tsunami survey on the island, Leonard and Bednarski (2015) collected a number of run-up observations ranging from ~ 3 to 13 m, peaking in the Davidson inlet in the near field of the tsunami source. The maximum inundation distances are relatively short due to steep coastal morphology; the tsunami penetrated the Pacific coast up to ~ 60 m, with maximum inundation distance

observed in the northernmost Otard Bay, some 130 km northwest from the epicenter.

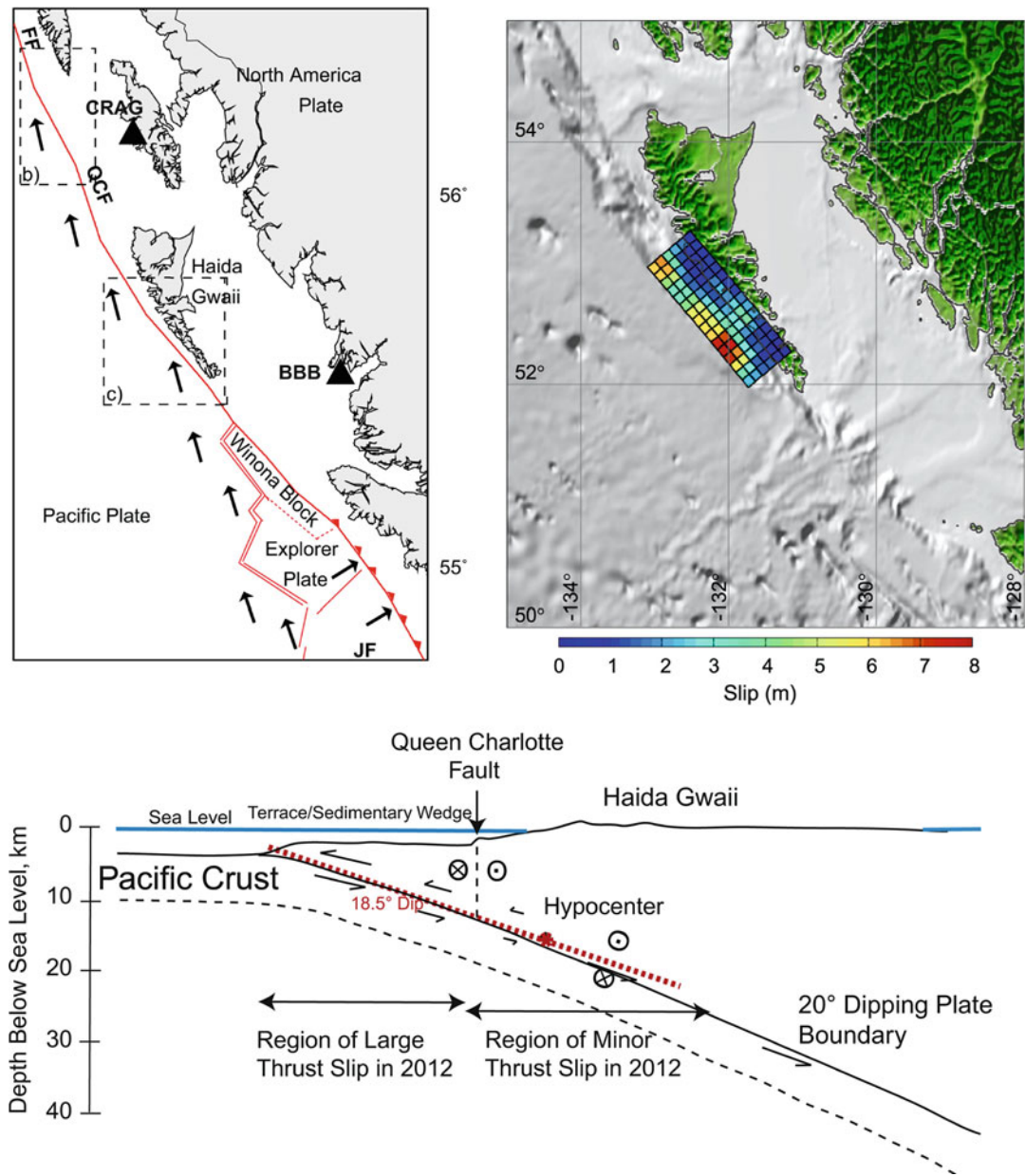
The tsunami was also detected by several tide gauges, as documented by the NGDC/WDS database and by Fine et al. (2015). It was recorded by several tide gauges located along the coast of British Columbia, in most cases with very small heights (< 20 cm). The highest tsunami waves (~ 0.5 m peak to trough) were recorded by the tide gauges in Henslung Cove on Langara Island (a small island north of Graham Island), and in Winter Harbour on the northwestern coast of Vancouver Island (Fine et al. 2015).

The tsunami also propagated over the northern Pacific Ocean; decimetric maximum heights were recorded by coastal tide gauges in the Hawaiian archipelago (at Kahului, the tide gauge recorded ~ 1.5 m from peak to trough); centimetric heights were measured elsewhere. Tsunami signals of several centimeters recorded by the DART buoys located off the Alaska, Oregon, Washington, and Canada coasts were used for the inference of the tsunami source, as described below. Additional tsunami observations arise from the ocean bottom pressure sensors of the NEPTUNE-Canada cabled observatory (Fine et al. 2015). Overall, in the far field, the Haida Gwaii earthquake generated modest tsunami waves. In terms of near-field effects, this was the largest tsunami ever recorded on Moresby Island.

Teleseismic data inversion, iteratively optimized by forward modeling of the tsunami signal at the surrounding DART buoys (Lay et al. 2013b, Fig. 24), indicates that the rupture started approximately beneath the trace of the Queen's Charlotte fault offshore of Moresby Island, but the coseismic slip release occurred essentially seaward of the hypocenter. The rupture propagated bilaterally along the trench direction for a total extent of ~ 150 km, extending seaward (up-dip) to the toe of the sedimentary wedge. GPS

Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 22 Triple junction and slip distribution (*top* panel, Lay 2015) (Reprinted from Earth and Planetary Science Letters, Vol.

409, Lay T., The surge of great earthquakes from 2004 to 2014, pp. 133–146, Copyright 2015, with permission from Elsevier) and run-up distribution (*bottom* panel, from Fritz et al. 2008) of the 2007 Solomon earthquake

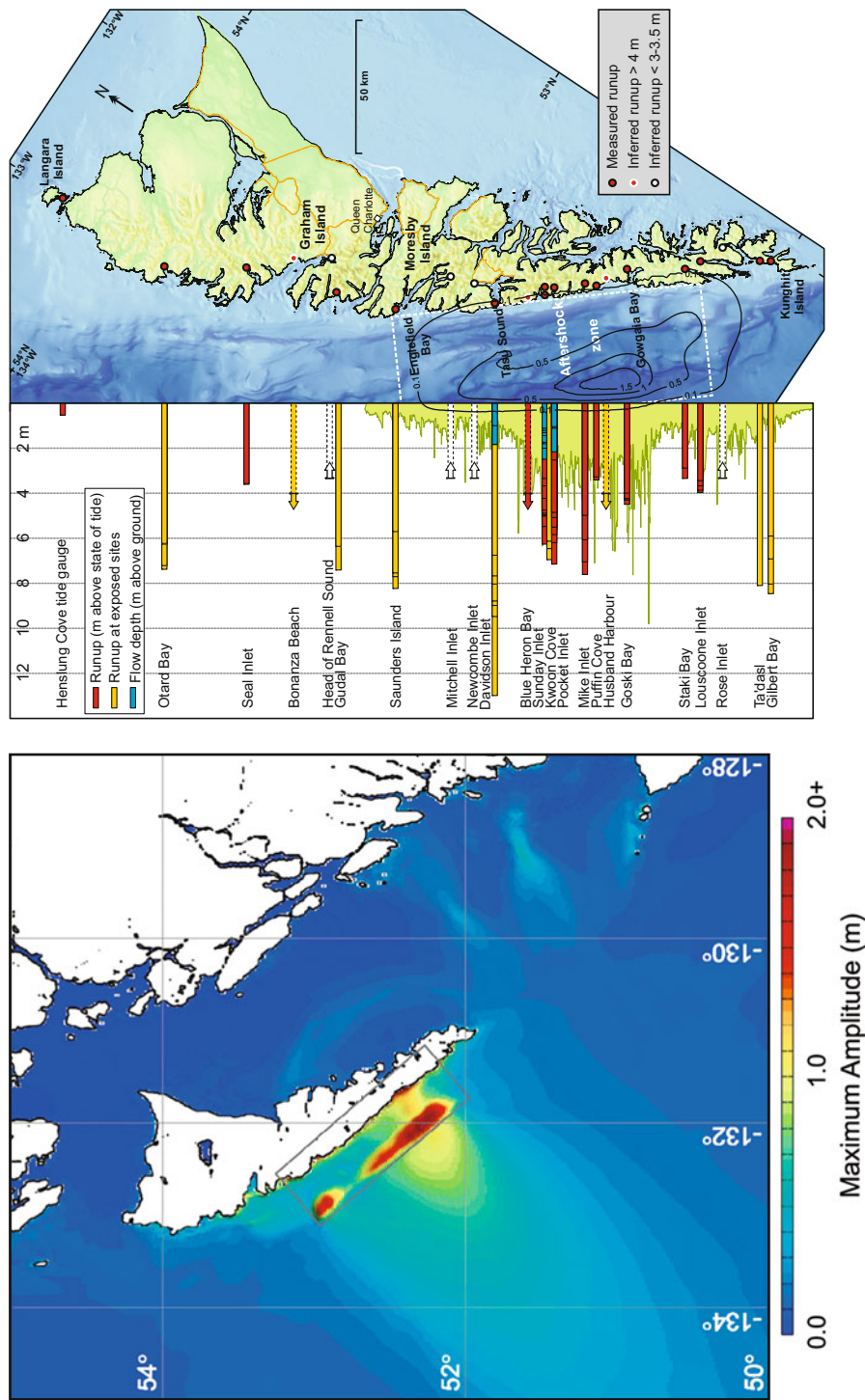


Tsunamigenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 23 Slip distribution and tectonic context/model of the 2012 Haida Gwaii earthquake (from Lay et al. 2013b) (Reprinted from Earth and Planetary Science Letters, Vol.

375, Lay T, Ye L, Kanamori H, Yamazaki Y, Cheung K F, Kwong K, Koper K D, Mw 7.8 Haida Gwaii underthrusting earthquake and tsunami: Slip partitioning along the Queen Charlotte Fault transpressional plate boundary, pp. 57-70, Copyright 2013, with permission from Elsevier)

observations indicate that the model of Lay et al. (2013b) has rupture extending too far down-dip beneath the coast, and a revised model with all slip locating offshore is preferred

(Nykolaishen et al. 2015). As a result, it appears that the slip was concentrated in a narrow band ~30 km wide on the subduction interface extending to the trench and localized entirely



Tsunamiogenic Major and Great Earthquakes (2004–2013): Source Processes Inverted from Seismic, Geodetic, and Sea-Level Data, Fig. 24 Maximum tsunami wave amplitudes (*left* panel, from Lay et al. 2013b) (Reprinted from Earth and Planetary Science Letters, Vol. 375, Lay T, Ye L, Kanamori H, Yamazaki Y, Cheung K F, Kwong K, Koper K D, Mw 7.8 Haida Gwaii underthrusting earthquake and tsunami: Slip partitioning along the Queen Charlotte Fault transpressional plate boundary, pp. 57–70,

Copyright 2013, with permission from Elsevier) and run-up distribution of the 2012 Haida Gwaii earthquake (*right* panel, from Fine et al. 2015) (Reprinted with kind permission from Springer Science+Business Media: Pure and Applied Geophysics, Observations and numerical modeling of the 2012 Haida Gwaii tsunami off the coast of British Columbia, vol.172, 2015, pp.699–718, Fine I V, Charniawsky J Y, Thomson R E, Rabinovich A B, Krassovski M V, Figure 14)

under the sedimentary wedge seaward of the Queen Charlotte Fault offshore of Moresby Island (Fig. 24).

As already noted for the 2006 Java and the 2010 Mentawai “tsunami earthquakes,” it is often assumed that the sedimentary wedge near the toe and the subduction interface beneath can only slip and deform aseismically. Conversely, this earthquake indicates that the interface below the Queen Charlotte terrace accumulates compressional strain perpendicular to the plate boundary and experiences stick-slip faulting. A surprising attribute of the faulting is that the rupture velocity inferred from seismic modeling is not particularly low (Lay et al. 2013b, find a rupture expansion velocity of 2.3 km/s), despite the shallow depth and sedimentary wedge contact zone. This is similar to the 2007 Solomon Islands earthquake, which ruptured the very shallow megathrust with comparable rupture expansion velocity (2.5 km/s). The moment-scaled radiated energy estimated by Lay et al. (2013b) is 1.28×10^{-5} , which is also higher than typical tsunami earthquakes, and the source spectrum is only slightly depleted in short-period spectral amplitudes. Thus, the Haida Gwaii earthquake is not a tsunami earthquake *per se*; this may be associated with the specific sedimentary materials, largely derived from erosion of the ancient metamorphic terrane uplifted in the islands, in contrast to deep abyssal muds on the plate interface in other trenches.

To first order, the largest run-up observations (Fig. 24) are spatially correlated with the tsunami source resulting from the teleseismic inversion (Leonard and Bednarski 2015). They lie mainly broadside to the inferred coseismic slip, with values ranging from ~ 3 to 13 m. In particular, large run-up values (~ 5 m), although not the largest, were observed in front of the broadest (along dip direction) slip patch, which also had the peak slip. They occurred inside Goski Bay, near the maximum (simulated) tsunami amplitudes, offshore the Pacific coast. The relatively flat area inside Goski Bay also features the maximum observed inundation distances of ~ 50 m. Notably, despite having occurred along a region of lower slip and smaller offshore tsunami

amplitudes, the absolute run-up maxima (~ 8 –13 m) were observed northernmost in the Davidson Inlet, with a shorter inundation distance than at other places (~ 35 m).

The observations indicate a complex relationship between the main source features, i.e., size and location of the slip patches, and the features of the generated tsunami, i.e., amplitude and period, which in turn determine the impact along the coastline. First, the slip pattern determines the amplitude and the steepness of the initial tsunami wave elevation, for example, the narrower the slip distribution, the shorter the tsunami wavelengths at the source and the shorter the period of the resulting waves. Second, the offshore tsunami amplitudes tend to peak near the broadest slip region, not necessarily in correspondence with the steepest and highest initial tsunami wave field. Third, an important role is played by the interaction of the tsunami period with the coastal features, with, for example, funneling and resonant phenomena causing higher run-up values for the short-period waves within the fjord-like Davidson Inlet (Leonard and Bednarski 2015), and longer inundation distances with run-up heights limited by frictional dissipation within Goski Bay for the longer-period tsunami components. This situation is similar to that already discussed for the Tohoku 2011 tsunami, with a correspondence between the tsunami periods generated in different source regions and the different inundation features observed in the Sendai plain and in the bays of the northernmost Iwate prefecture.

Another interesting feature is the very sustained inundation observed at Gilbert Bay, peaking at ~ 9 m and reaching to ~ 50 m from the coast, occurring in the near-field oblique regime with respect to the source. This is likely caused by the late arrival of trapped edge waves, which traveled along the coast of Moresby Island.

Future Directions

The decade from 2004 to 2013 is striking both for the number of major tsunamigenic events and the

great damage and loss of life produced by tsunami. It is also notable for the vastly expanded seismic, geodetic, and tsunami recordings available to analyze the source processes, tsunami excitation, and complexity of the events. There are still major limitations in instrument coverage, and not all events have comparable recordings, but instrument deployments are steadily increasing, and operational earthquake and tsunami warning capabilities are advancing in many regions.

Progress in quantification of the nature of great earthquake ruptures has been dramatic, but the recent events also exhibit great complexity in individual ruptures and in triggering interactions that present many challenges. Compound faulting and doublet events are very challenging for warning systems, as are shallow “tsunami earthquakes.” The analyses of multiple datasets have demonstrated the complementary sensitivity to source processes provided by seismic, geodetic, and tsunami observation and the value of joint analysis when possible. Performing such comprehensive analyses in near real time to augment tsunami warning systems is technically possible but remains very challenging logistically.

Improved assessment of strain accumulation distributions, using both onshore and offshore geodetic methods, is very desirable for identifying regions of likely future large-slip events and for making progress on bounding likely rupture extent beyond the limited performance of classic seismic gap model approaches. Tsunami hazard assessments need to consider not only ordinary interplate megathrust faulting but also “extremes” in which shallow slip is enhanced and even “tsunami earthquakes” or intraplate rupture possibilities in many regions. Nonetheless, probabilities for these features are difficult to assess. Great caution is moreover needed regarding assumptions on earthquake segmentation, and the possibility that an M9 earthquake may occur virtually on any subduction zone should not be overlooked.

Acknowledgments The authors are grateful to Eric L. Geist for his constructive comments, to the Section Editors Eric L. Geist and W. H. K. Lee, and to the Senior Editor Andrew Spencer. We acknowledge large use of sea-level and tsunami data from the following websites: <https://www.ngdc.noaa.gov/hazard/tide.shtml>; <http://nctr.pmel.noaa.gov/Dart/>;

<http://www.ioc-sealevelmonitoring.org>; and of earthquake parametric data and information from <http://earthquake.usgs.gov/contactus/golden/neic.php> and other USGS web pages. Global seismic recordings for the analyses discussed here were primarily obtained from the data centers of the Federation of Digital Seismic Networks and the Incorporated Research Institutions for Seismology. We also made large use of the International Tsunami Information Center website (<http://itic.ioc-unesco.org>). Some figures were produced with Generic Mapping Tools (GMT, <https://www.soest.hawaii.edu/gmt/>). We are grateful to the numerous colleagues who sent us high-resolution versions of all the figures reproduced in this paper (with permission from publisher when requested). S. Lorito's and F. Romano's work on earthquakes and tsunamis is supported by Italian Flagship project RITMARE; by project ASTARTE (Assessment, Strategy And Risk Reduction for Tsunamis in Europe) FP7-ENV2013 6.4-3, grant 603839; by the agreement between Italian DPC and INGV, annex B2 (2015). T. Lay's work on earthquakes is supported by NSF grant EAR1245717.

Bibliography

- Ammon CJ, Ji C, Thio HK, Robinson D, Ni S, Hjorleifsdottir V, Kanamori H, Lay T, Das S, Helmberger D, Ichinose G, Polet J, Wald D (2005) Rupture process of the 2004 Sumatra-Andaman earthquake. *Science* 308:1133–1139. <https://doi.org/10.1126/science.1112260>
- Ammon CJ, Kanamori H, Lay T, Velasco AA (2006) The 17 July 2006 Java tsunami earthquake. *Geophys Res Lett* 33:L24308. <https://doi.org/10.1029/2006GL028005>
- Ammon CJ, Kanamori H, Lay T (2008) A great earthquake doublet and seismic stress transfer cycles in the Central Kuril Islands. *Nature* 451:561–565. <https://doi.org/10.1038/nature06521>
- Ammon CJ, Lay T, Simpson DW (2010) Great earthquakes and global seismic networks. *Seismol Res Lett* 81:965–971. <https://doi.org/10.1785/gssrl.81.6.965>
- Baba T, Cummins PR, Thio HK, Tsushima H (2009) Validation and joint inversion of teleseismic waveforms for earthquake source models using deep ocean bottom pressure records: a case study of the 2006 Kuril megathrust earthquake. *Pure Appl Geophys* 166:55–76. <https://doi.org/10.1007/s00024-008-0438-1>
- Banerjee P, Pollitz FF, Nagarajan B, Bürgmann R (2007) Coseismic slip distributions of the 26 December 2004 Sumatra-Andaman and 28 March 2005 Nias earthquakes from GPS static offsets. *Bull Seismol Soc Am* 97:S86–S102. <https://doi.org/10.1785/0120050609>
- Beavan J, Wang X, Holden C, Wilson K, Power W, Prasetya G, Bevis M, Kautoke R (2010) Near-simultaneous great earthquakes at Tongan megathrust and outer rise in September 2009. *Nature* 466:959–963. <https://doi.org/10.1038/nature09292>

- Beck S, Barrientos S, Kausel E, Reyes M (1998) Source characteristics of historic earthquakes along the central Chile subduction zone. *J South Am Earth Sci* 11:115–129
- Biggs J, Robinson DP, Dixon TH (2009) The 2007 Pisco, Peru, earthquake (M 8.0): seismology and geodesy. *Geophys J Int* 176:657–669. <https://doi.org/10.1111/j.1365-246X.2008.03990.x>
- Bilham R (2005) A flying start, then a slow slip. *Science* 308:1126–1127. <https://doi.org/10.1126/science.1113363>
- Bletery Q, Sladen A, Delouis B, Vallée M, Nocquet JM, Rolland L, Jiang J (2014) A detailed source model for the Mw 9.0 Tohoku-Oki earthquake reconciling geodesy, seismology, and tsunami records. *J Geophys Res* 119:7636–7653. <https://doi.org/10.1002/2014JB011261>
- Borrero JC, Weiss R, Okal EA, Hidayat R, Arcas Suranto D, Titov VV (2009) The tsunami of 2007 September 12, Bengkulu province, Sumatra, Indonesia: post-tsunami field survey and numerical modelling. *Geophys J Int* 178:180–194. <https://doi.org/10.1111/j.1365-246X.2008.04058.x>
- Borrero JC, McAdoo B, Jaffe B, Dengler L, Gelfenbaum G, Higman B, Hidayat R, Moore A, Kongko W, Lukijanto PR, Prasetya G, Titov V, Yulianto E (2012) Field survey of the March 28, 2005 Nias-Simeulue earthquake and tsunami. *Pure Appl Geophys* 168:1075–1088. <https://doi.org/10.1007/s00024-010-0218-6>
- Briggs RW, Sieh K, Meltzner AJ, Natawidjaja D, Galetzka J, Suwargadi B, Hsu YJ, Simons M, Hananto N, Suprihanto I, Prayudi D, Avouac JP, Prawirodirdjo L, Bock Y (2006) Deformation and slip along the Sunda megathrust in the great 2005 Nias-Simeulue earthquake. *Science* 311:1897–1901. <https://doi.org/10.1126/science.1122602>
- Chen T, Newman AV, Feng L, Fritz HM (2009) Slip distribution from the 1 April 2007 Solomon Islands earthquake: a unique image of the near-trench rupture. *Geophys Res Lett* 36:L16307. <https://doi.org/10.1029/2009GL039496>
- Chlieh M, Avouac JP, Hjorleifsdottir V, Song TR, Ji C, Sieh K, Sladen A, Hebert H, Prawirodirdjo L, Bock Y, Galetzka J (2007) Coseismic slip and afterslip of the great (Mw 9.15) Sumatra-Andaman earthquake of 2004. *Bull Seismol Soc Am* 97:S152–S173. <https://doi.org/10.1785/0120050631>
- Choi BH, Hong SJ, Pelinovsky E (2006) Distribution of runup heights of the December 26, 2004 tsunami in the Indian Ocean. *Geophys Res Lett* 33:L13601. <https://doi.org/10.1029/2006GL025867>
- Das S (2011) Earthquake rupture: inverse problem. In: Gupta HK (ed) *Encyclopedia of solid earth geophysics*. Springer, Netherlands. <https://doi.org/10.1007/978-90-481-8702-7>
- Delescluse M, Chamot-Rooke N, Cattin R, Fleitout L, Trubienko O, Vigny C (2012) April 2012 intra-oceanic seismicity off Sumatra boosted by the Banda-Aceh megathrust. *Nature* 490:240–244. <https://doi.org/10.1038/nature11520>
- Delouis B, Nocquet JM, Vallée M (2010) Slip distribution of the February 27, 2010 Mw = 8.8 Maule earthquake, central Chile, from static and high-rate GPS, InSAR, and broadband teleseismic data. *Geophys Res Lett* 37:L17305. <https://doi.org/10.1029/2010GL043899>
- Eberhart-Phillips D, Haeussler PJ, Freymueller JT, Frankel AD, Rubin CM, Craw P, Ratchkovski NA, Anderson G, Carver GA, Crone AJ et al (2003) The 2002 Denali fault earthquake, Alaska: a large magnitude, slip-partitioned event. *Science* 300:1113–1118. <https://doi.org/10.1126/science.1082703>
- Faulkner DR, Mitchell TM, Behn J, Hirose T, Shimamoto T (2011) Stuck in the mud? Earthquake nucleation and propagation through accretionary forearcs. *Geophys Res Lett* 38:L18303. <https://doi.org/10.1029/2011GL048552>
- Fine IV, Cherniawsky JY, Thomson RE, Rabinovich AB, Krassovski MV (2015) Observations and numerical modeling of the 2012 Haida Gwaii tsunami off the coast of British Columbia. *Pure Appl Geophys* 172:699–718. <https://doi.org/10.1007/s00024-014-1012-7>
- Fritz HM, Kalligeris N (2008) Ancestral heritage saves tribes during 1 April 2007 Solomon Islands tsunami. *Geophys Res Lett* 35:L01607. <https://doi.org/10.1029/2007GL031654>
- Fritz HM, Kongko W, Moore A, McAdoo B, Goff J, Harbitz C, Uslu B, Kalligeris N, Suteja D, Kalsum K, Titov V, Gusman A, Latief H, Santoso E, Sujoko S, Djulkarnaen D, Sunendar H, Synolakis C (2007) Extreme runup from the 17 July 2006 Java tsunami. *Geophys Res Lett* 34:L12602. <https://doi.org/10.1029/2007GL029404>
- Fritz HM, Kalligeris N, Borrero JC, Broncano P, Ortega E (2008) The 15 August 2007 Peru tsunami runup observations and modeling. *Geophys Res Lett* 35:L10604. <https://doi.org/10.1029/2008GL033494>
- Fritz HM, Borrero JC, Synolakis CE, Okal EA, Weiss R, Titov VV, Jaffe BE, Foteinis S, Lynett PJ, Chan I, Liu PLF (2011a) Insights on the 2009 South Pacific tsunami in Samoa and Tonga from field surveys and numerical simulations. *Earth Sci Rev* 107:66–75. <https://doi.org/10.1016/j.earscirev.2011.03.004>
- Fritz HM, Petroff CM, Catalán CR, Winckler P, Kalligeris N, Weiss R (2011b) Field survey of the 27 February 2010 Chile tsunami. *Pure Appl Geophys* 168:1989–2010. <https://doi.org/10.1007/s00024-011-0283-5>
- Fritz H M, Papantoniou A, Biukoto L, Gilly A, Wei Y (2014) The Solomon Islands Tsunami of 6 February 2013 in the Santa Cruz Islands: Field Survey and Modeling, EGU General Assembly 2014, held 27 April - 2 May, in Vienna, Austria, id.15777
- Fujii Y, Satake K (2006) Source of the July 2006 West Java tsunami estimated from tide gauge records. *Geophys Res Lett* 33:L24317. <https://doi.org/10.1029/2006GL028049>

- Fujii Y, Satake K (2007) Tsunami source of the 2004 Sumatra-Andaman earthquake inferred from tide gauge and satellite data. *Bull Seismol Soc Am* 97: S192–S207. <https://doi.org/10.1785/0120050613>
- Fujii Y, Satake K (2008) Tsunami sources of the November 2006 and January 2007 Great Kuril earthquakes. *Bull Seismol Soc Am* 98:1559–1571. <https://doi.org/10.1785/0120070221>
- Fulton PM, Brodsky EE, Kano Y, Mori J, Chester F, Ishikawa T, Harris RN, Lin W, Eguchi N, Toczko S, Expedition 343, 343T, and KR13-08 Scientists (2013) Low coseismic friction on the Tohoku-Oki fault determined from temperature measurements. *Science* 342:1214–1217. <https://doi.org/10.1126/science.1243641>
- Furlong KP, Lay T, Ammon CJ (2009) A great earthquake rupture across a rapidly evolving three-plate boundary. *Science* 324:226–229. <https://doi.org/10.1126/science.1167476>
- Gahalaut VK, Nagarajan B, Catherine JK, Kumar S (2006) Constraints on 2004 Sumatra-Andaman earthquake rupture from GPS measurements in Andaman-Nicobar islands. *Earth Planet Sci Lett* 242:365–374. <https://doi.org/10.1016/j.epsl.2005.11.051>
- Geist EL (2009) Phenomenology of tsunamis: statistical properties from generation to runup. *Adv Geophys* 51:107–169. [https://doi.org/10.1016/S0065-2687\(09\)05108-5](https://doi.org/10.1016/S0065-2687(09)05108-5)
- Geist EL (2012) Near-field tsunami edge waves and complex earthquake rupture. *Pure Appl Geophys* 170:1475–1491
- Geist EL, Bilek SL (2001) Effect of depth-dependent shear modulus on tsunami generation along subduction zones. *Geophys Res Lett* 28:1315–1318
- Geist EL, Oglesby DD (2014) Tsunamis: stochastic models of occurrence and generation mechanisms. In: Robert A. Meyers (ed.) *Encyclopedia of complexity and systems science*, Springer, New York. https://doi.org/10.1007/978-3-642-27737-5_595-1
- Geist EL, Parsons T (2005) Triggering of tsunamigenic aftershocks from large strike-slip earthquakes: analysis of the November 2000 New Ireland earthquake sequence. *Geochem Geophys Geosyst* 6:Q10005. <https://doi.org/10.1029/2005GC000935>
- Geist EL, Parsons T (2011) Assessing historical rate changes in global tsunami occurrence. *Geophys J Int* 187:497–509. <https://doi.org/10.1111/j.1365-246X.2011.05160.x>
- Geist EL, Bilek SL, Arcas D, Titov VV (2006) Differences in tsunami generation between the December 26, 2004 and March 28, 2005 Sumatra earthquakes. *Earth Planets Space* 58:185–193. <https://doi.org/10.1186/BF03353377>
- González FI et al (2009) Probabilistic tsunami hazard assessment at Sea-side, Oregon, for near- and far-field seismic sources. *J Geophys Res* 114:C11023. <https://doi.org/10.1029/2008JC005132>
- Gusman AR, Tanioka Y, Kobayashi T, Latief H, Pandoe W (2010) Slip distribution of the 2007 Bengkulu earthquake inferred from tsunami waveforms and InSAR data. *J Geophys Res* 115:B12316. <https://doi.org/10.1029/2010JB007565>
- Hartzell SH, Heaton TH (1983) Inversion of strong ground motion and teleseismic waveform data for the fault rupture history of the 1979 Imperial Valley, California, earthquake. *Bull Seismol Soc Am* 73:1553–1583
- Hayes GP, Furlong KP, Benz HM, Herman HW (2014) Triggered aseismic slip adjacent to the 6 February 2013 Mw8.0 Santa Cruz Islands megathrust earthquake. *Earth Planet Sci Lett* 388:265–272. <https://doi.org/10.1016/j.epsl.2013.11.010>
- Hill EM, Borrero JC, Huang Z, Qiu Q, Banerjee P, Natawidjaja DH, Elosegui P, Fritz HM, Suwargadi BW, Pranantyo IR, Li LL, Macpherson KA, Skanavis V, Synolakis CE, Sieh K (2012) The 2010 Mw 7.8 Mentawai earthquake: very shallow source of a rare tsunami earthquake determined from tsunami field survey and near-field GPS data. *J Geophys Res* 117: B06402. <https://doi.org/10.1029/2012JB009159>
- Hill EM, Yue H, Barbot S, Lay T, Tapponnier P, Hermawan I, Hubbard J, Banerjee P, Feng L, Natawidjaja D, Sieh K (2015) The 2012 Mw 8.6 Wharton Basin sequence: a cascade of great earthquakes generated by near-orthogonal, young, oceanic-mantle faults. *J Geophys Res* 120:1–25. <https://doi.org/10.1002/2014JB011703>
- Hoechner A, Babeyko AY, Sobolev SV (2008) Enhanced GPS inversion technique applied to the 2004 Sumatra earthquake and tsunami. *Geophys Res Lett* 35:L08310. <https://doi.org/10.1029/2007GL033133>
- Hossen MJ, Cummins PR, Dettmer J, Baba T (2015) Tsunami waveform inversion for sea surface displacement following the 2011 Tohoku earthquake: importance of dispersion and source kinematics. *J Geophys Res Solid Earth* 120:1–22. <https://doi.org/10.1002/2015JB011942>
- Hsu YJ, Simons M, Avouac JP, Galetzka J, Sieh K, Chlieh M, Natawidjaja D, Prawirodirdjo L, Bock Y (2006) Frictional afterslip following the 2005 Nias-Simeulue earthquake, Sumatra. *Science* 312:1921–1925. <https://doi.org/10.1126/science.1126960>
- Hsu YJ, Simons M, Williams C, Casarotti E (2011) Three-dimensional FEM derived elastic Green's functions for the coseismic deformation of the 2005 Mw 8.7 Nias-Simeulue, Sumatra earthquake. *Geochem Geophys Geosyst* 12:1–19. <https://doi.org/10.1029/2011GC003553>
- Hughes K, Masterlark T, Mooney WD (2010) Poroelastic stress-triggering of the 2005 Mw8.7 Nias earthquake by the 2004 Mw9.2 Sumatra-Andaman earthquake. *Earth Planet Sci Lett* 293:289–299. <https://doi.org/10.1016/j.epsl.2010.02.043>
- Ioualalen M, Perfettini H, Condo SY, Jimenez C, Tavera H (2013) Tsunami modeling to validate slip models of the 2007 Mw8.0 Pisco earthquake, Central Peru. *Pure Appl Geophys* 170:433–451. <https://doi.org/10.1007/s00024-012-0608-z>
- Ishii M, Shearer PM, Houston H, Vidale JE (2007) Teleseismic P wave imaging of the 26 December

- 2004 Sumatra-Andaman and 28 March 2005 Sumatra earthquake ruptures using the Hi-net array. *J Geophys Res* 112:B11307. <https://doi.org/10.1029/2006JB004700>
- Ishii M, Kiser E, Geist EL (2012) Mw 8.6 Sumatran earthquake of 11 April 2012: rare seaward expression of oblique subduction. *Geology* 41:319–322. <https://doi.org/10.1130/G33783.1>
- Jackson DD, Kagan YY (2011) Characteristic earthquakes and seismic gaps. In: Harsh K. Gupta (ed.) *Encyclopedia of solid earth geophysics*
- Kagan YY, Jackson DD (2013) Tohoku earthquake: a surprise? *Bull Seismol Soc Am* 103:1181–1194. <https://doi.org/10.1785/0120120110>
- Kanamori H (1971) Seismological evidence for a lithospheric normal faulting – the Sanriku earthquake of 1933. *Phys Earth Planet In* 4:289–300
- Kanamori H (1972) Mechanism of tsunami earthquakes. *Phys Earth Planet Inter* 6:346–359. [https://doi.org/10.1016/0031-9201\(72\)90058-1](https://doi.org/10.1016/0031-9201(72)90058-1)
- Kanamori H (1977) The energy release of great earthquakes. *J Geophys Res* 82:2981–2987
- Kanamori H, McNally KC (1982) Variable rupture mode of the subduction zone along the Ecuador-Colombia coast. *Bull Seismol Soc Am* 72:1241–1253
- Kanamori H, Rivera L, Lee WHK (2010) Historical seismograms for unraveling a mysterious earthquake: the 1907 Sumatra earthquake. *Geophys J Int* 183:358–374. <https://doi.org/10.1111/j.1365-246X.2010.04731.x>
- Kaneko Y, Avouac JP, Lapusta N (2010) Towards inferring earthquake patterns from geodetic observations of interseismic coupling. *Nat Geosci* 3:363–369. <https://doi.org/10.1038/NGEO843>
- Kánoğlu U, Titov VV, Aydın B, Moore C, Stefanakis TS, Zhou H, Spillane M, Synolakis CE (2013) Focusing of long waves with finite crest over constant depth. *Proc R Soc A* 469:1–20. <https://doi.org/10.1098/rspa.2013.0015>
- Konca AO, Hjørleifsdóttir V, Song TRA, Avouac JP, Helmberger DV, Ji C, Sieh K, Briggs R, Meltzner A (2007) Rupture kinematics of the 2005 M_w 8.6 Nias-Simeulue earthquake from the joint inversion of seismic and geodetic data. *Bull Seismol Soc Am* 97: S307–S322. <https://doi.org/10.1785/0120050632>
- Konca AO, Avouac JP, Sladen A, Meltzner AJ, Sieh K, Fang P, Li Z, Galetzka J, Genrich J, Chlieh M, Natawidjaja DH, Bock Y, Fielding EJ, Ji C, Helmberger DV (2008) Partial rupture of a locked patch of the Sumatra megathrust during the 2007 earthquake sequence. *Nature* 456:631–635. <https://doi.org/10.1038/nature07572>
- Koper KD, Hutko AR, Lay T, Sufri O (2012) Imaging short-period seismic radiation from the 27 February 2010 Chile (Mw 8.8) earthquake by back-projection of P, PP, and PKIKP waves. *J Geophys Res* 117: B02308. <https://doi.org/10.1029/2011JB008576>
- Kowalik Z, Knorrillo J, Knorrillo W, Logan T (2008) Kuril Islands tsunami of November 2006: 1. Impact at Crescent City by distant scattering. *J Geophys Res* 113:C01020. <https://doi.org/10.1029/2007JC004402>
- Kozdon JE, Dunham EM (2013) Rupture to the trench: dynamic rupture simulations of the 11 March 2011 Tohoku earthquake. *Bull Seismol Soc Am* 103:1275–1289. <https://doi.org/10.1785/0120120136>
- Kubo H, Kakehi Y (2013) Source process of the 2011 Tohoku earthquake estimated from the joint inversion of teleseismic body waves and geodetic data including seafloor observation data: source model with enhanced reliability by using objectively determined inversion settings. *Bull Seismol Soc Am* 103:1195–1220. <https://doi.org/10.1785/0120120113>
- Lay T (2015) The surge of great earthquakes from 2004 to 2014. *Earth Planet Sci Lett Invited Front Pap* 409:133–146. <https://doi.org/10.1016/j.epsl.2014.10.047>
- Lay T, Kanamori H (1980) Earthquake doublets in the Solomon Islands. *Phys Earth Planet In* 21:283–304
- Lay T, Kanamori H, Ammon CJ, Nettles M, Ward SN, Aster RC, Beck SL, Bilek SL, Brudzinski MR, Butler R, DeShon HR, Ekström G, Satake K, Sipkin S (2005a) The Great Sumatra-Andaman earthquake of the 26 December 2004. *Science* 308:1127–1133. <https://doi.org/10.1126/science.1112250>
- Lay T, Kanamori H, Ammon CJ, Nettles M, Ward SN, Aster RC, Beck SL, Bilek SL, Brudzinski MR, Butler R, DeShon HR, Ekström G, Satake K, Sipkin S (2005b) Response to comment on “The Great Sumatra-Andaman Earthquake of the 26 December 2004”. *Science* 310:1431. <https://doi.org/10.1126/science.1119662>
- Lay T, Kanamori H, Ammon CJ, Hutko AR, Furlong K, Rivera L (2009) The 2006–2007 Kuril Islands great earthquake sequence. *J Geophys Res* 114:B113208. <https://doi.org/10.1029/2008JB006280>
- Lay T, Ammon CJ, Kanamori H, Koper KD, Sufri O, Hutko AR (2010a) Teleseismic inversion for rupture process of the 27 February 2010 Chile (Mw 8.8) earthquake. *Geophys Res Lett* 37:L13301. <https://doi.org/10.1029/2010GL043379>
- Lay T, Ammon CJ, Hutko AR, Kanamori H (2010b) Effects of kinematic constraints on teleseismic finite-source rupture inversions: Great Peruvian earthquakes of 23 June 2001 and 15 August 2007. *Bull Seismol Soc Am* 100:969–994. <https://doi.org/10.1785/0120090274>
- Lay T, Ammon CJ, Kanamori H, Rivera L, Koper KD, Hutko AR (2010c) The 2009 Samoa–Tonga great earthquake triggered doublet. *Nature* 466:964–968. <https://doi.org/10.1038/nature09214>
- Lay T, Ammon CJ, Kanamori H, Kim MJ, Xue L (2011a) Outer trench-slope faulting and the 2011 Mw 9.0 off the Pacific coast of Tohoku earthquake. *Earth Planets Space* 63:713–718. <https://doi.org/10.5047/eps.2011.05.006.713-718>
- Lay T, Ammon CJ, Kanamori H, Xue L, Kim MJ (2011b) Possible large near-trench slip during the 2011 Mw 9.0 off the Pacific coast of Tohoku earthquake. *Earth*

- Planets Space 63:687–692. <https://doi.org/10.5047/eps.2011.05.033>
- Lay T, Ammon CJ, Kanamori H, Yamazaki Y, Cheung KF, Hutko AR (2011c) The 25 October 2010 Mentawai tsunami earthquake (Mw 7.8) and the tsunami hazard presented by shallow megathrust ruptures. *Geophys Res Lett* 38:L06302. <https://doi.org/10.1029/2010GL046552>
- Lay T, Kanamori H, Ammon CJ, Koper KD, Hutko AR, Ye L, Yue H, Rushing TM (2012) Depth-varying rupture properties of subduction zone megathrust faults. *J Geophys Res* 117:B04311. <https://doi.org/10.1029/2011JB009133>
- Lay T, Ye L, Kanamori H, Yamazaki Y, Cheung KF, Ammon CJ (2013a) The February 6, 2013 Mw 8.0 Santa Cruz Islands earthquake and tsunami. *Tectonophysics* 608:1109–1121. <https://doi.org/10.1016/j.tecto.2013.07.001>
- Lay T, Ye L, Kanamori H, Yamazaki Y, Cheung KF, Kwong K, Koper KD (2013b) Mw 7.8 Haida Gwaii underthrusting earthquake and tsunami: slip partitioning along the Queen Charlotte Fault transpressional plate boundary. *Earth Planet Sci Lett* 375:57–70. <https://doi.org/10.1016/j.epsl.2013.05.005>
- Leonard LJ, Bednarski JM (2015) The preservation potential of coastal coseismic and tsunami evidence observed following the 2012 Mw 7.8 Haida Gwaii Thrust earthquake. *Bull Seismol Soc Am* 105:1280–1289. <https://doi.org/10.1785/0120140193>
- Lin YN, Sladen A, Ortega-Culaciati F, Simons M, Avouac JP, Fielding EJ, Brooks BA, Bevis M, Genrich J, Rietbrock A, Vigny C, Smalley R, Socquet A (2013) Coseismic and postseismic slip associated with the 2010 Maule earthquake, Chile: characterizing the Arauco Peninsula barrier effect. *J Geophys Res* 118:3142–3159. <https://doi.org/10.1002/jgrb.50207>
- Liu PLF, Yeh H, Lin P, Chang KT, Cho YS (1998) Generation and evolution of edge-wave packets. *Phys Fluids* 10:1635–1657
- Lorito S, Romano F, Piatanesi A, Boschi E (2008) Source process of the September 12, 2007, Mw 8.4 southern Sumatra earthquake from tsunami tide gauge record inversion. *Geophys Res Lett* 35:L02310. <https://doi.org/10.1029/2007GL032661>
- Lorito S, Piatanesi A, Cannelli V, Romano F, Melini D (2010) Kinematics and source zone properties of the 2004 Sumatra-Andaman earthquake and tsunami: nonlinear joint inversion of tide-gauge, satellite altimetry, and GPS data. *J Geophys Res* 115:B02304. <https://doi.org/10.1029/2008JB005974>
- Lorito S, Romano F, Atzori S, Tong X, Avallone A, McCloskey J, Cocco M, Boschi E, Piatanesi A (2011) Limited overlap between the seismic gap and coseismic slip of the great 2010 Chile earthquake. *Nat Geosci* 4:173–177. <https://doi.org/10.1038/ngeo1073>
- Loveless JP, Meade BJ (2010) Geodetic imaging of plate motions, slip rates, and partitioning of deformation in Japan. *J Geophys Res* 115:B02410. <https://doi.org/10.1029/2008JB006248>
- Lynnes CS, Lay T (1988) Source process of the great 1977 Sumba earthquake. *J Geophys Res* 93:13407–13420
- MacInnes BT, Pinegina TK, Bourgeois J, Razhigaeva NG, Kaistrenko VM, Kravchunovskaya EA (2009) Field survey and geological effects of the 15 November 2006 Kuril Tsunami in the Middle Kuril Islands. *Pure Appl Geophys* 166:9–36. <https://doi.org/10.1007/s00024-008-0428-3>
- Madariaga R, Métois M, Vigny C, Campos J (2010) Central Chile finally breaks. *Science* 328:181–182. <https://doi.org/10.1126/science.1189197>
- Maercklin N, Festa G, Colombelli S, Zollo A (2012) Twin ruptures grew to build up the giant 2011 Tohoku, Japan, earthquake. *Sci Rep* 709:1–6. <https://doi.org/10.1038/srep00709>
- McCloskey J, Lange D, Tilmann F, Nalbant SS, Bell AF, Natawidjaja DH, Rietbrock A (2010) The September 2009 Padang earthquake. *Nat Geosci* 3:70–71. <https://doi.org/10.1038/ngeo753>
- Menke W, Abend H, Bach D, Newman K, Levin V (2006) Review of the source characteristics of the Great Sumatra-Andaman islands earthquake of 2004. *Surv Geophys* 27:603–613. <https://doi.org/10.1007/s10712-006-9013-4>
- Minson SE, Simons M, Beck JL (2013) Bayesian inversion for finite fault earthquake source models I – theory and algorithm. *Geophys J Int* 194:1701–1726. <https://doi.org/10.1093/gji/ggt180>
- Moreno M, Rosenau M, Oncken O (2010) 2010 Maule earthquake slip correlates with pre-seismic locking of Andean subduction zone. *Nature* 467:198–204. <https://doi.org/10.1038/nature09349>
- Moreno M, Melnick D, Rosenau M, Baez J, Klotz J, Oncken O, Tassara A, Chen J, Bataille K, Bevis M, Socquet A, Bolte J, Vigny C, Brooks B, Ryder I, Grund V, Smalley B, Carrizo D, Bartsch M, Hase H (2012) Toward understanding tectonic control on the Mw 8.8 2010 Maule Chile earthquake. *Earth Planet Sci Lett* 321–322:152–165. <https://doi.org/10.1016/j.epsl.2012.01.006>
- Mori N, Takahashi T, Yasuda T, Yanagisawa H (2011) Survey of the 2011 Tohoku earthquake tsunami inundation and run-up. *Geophys Res Lett* 38:L00G14. <https://doi.org/10.1029/2011GL049210>
- Munger S, Cheung KF (2008) Resonance in Hawaii waters from the 2006 Kuril Islands tsunami. *Geophys Res Lett* 35:L07605. <https://doi.org/10.1029/2007GL032843>
- Nalbant SS, Steacy S, Sieh K, Natawidjaja D, McCloskey J (2005) Earthquake risk on the Sunda trench. *Nature* 435:756–757. <https://doi.org/10.1038/nature435756a>
- Natawidjaja DH, Sieh K, Chlieh M, Galetzka J, Suwargadi BW, Cheng H, Edwards RL, Avouac JP, Ward SN (2006) Source parameters of the great Sumatran megathrust earthquakes of 1797 and 1833 inferred from coral microatolls. *J Geophys Res* 111:B06403. <https://doi.org/10.1029/2005JB004025>

- Neetu S, Suresh I, Shankar R, Shankar D, Shenoi SSC, Shetye SR, Sundar D, Nagarajan B (2005) Comment on "The Great Sumatra-Andaman Earthquake of the 26 December 2004". *Science* 310:1431. <https://doi.org/10.1126/science.1118950>
- Newman AV, Hayes G, Wei Y, Convers J (2011) The 25 October 2010 Mentawai tsunami earthquake, from real-time discriminants, finite-fault rupture, and tsunami excitation. *Geophys Res Lett* 38:L05302. <https://doi.org/10.1029/2010GL046498>
- Nishenko SP (1991) Circum-Pacific seismic potential: 1989–1999. *Pure Appl Geophys* 135:169–259
- Nykolaishen L, Dragert H, Wang K, James TS, Schmidt M (2015) GPS observations of crustal deformation associated with the 2012 Mw 7.8 Haida Gwaii earthquake. *Bull Seismol Soc Am* 105:1241–1252. <https://doi.org/10.1785/0120140177>
- Oglesby DD, Archuleta RJ, Nielsen SB (1998) Earthquakes on dipping faults: the effects of broken symmetry. *Science* 280:1055–1059. <https://doi.org/10.1126/science.280.5366.1055>
- Okal EA (1988) Seismic parameters controlling Far-field tsunami amplitudes: a review. *Nat Hazards* 1:67–96
- Okal EA, Fritz HM, Synolakis CE, Borrero JC, Weiss R, Lynett PJ, Titov VV, Foteinis S, Jaffe BE, Liu PLF, Chan I (2010) Field survey of the Samoa tsunami of 29 September 2009. *Seismol Res Lett* 81:577–591. <https://doi.org/10.1785/gssrl.81.4.577>
- Park J, Song T-R A, Tromp J, Okal E, Stein S, Roullet G, Clevede E, Laske G, Kanamori H, Davis P, Berger J, Braitenberg C, Van Camp M, Lei X, Sun H, Xu H, Rosat S (2005) Earth's free oscillations excited by the 26 December 2004 Sumatra-Andaman earthquake. *Science* 308(5725):1139–1144. <https://doi.org/10.1126/science.1112305>
- Philibosian B, Sieh K, Avouac JP, Natawidjaja DH, Chiang HW, Wu CC, Perfettini H, Shen CC, Daryono MR, Suwargadi BW (2014) Rupture and variable coupling behavior of the Mentawai segment of the Sunda megathrust during the supercycle culmination of 1797 to 1833. *J Geophys Res* 119:7258–7287. <https://doi.org/10.1002/2014JB011200>
- Piatanesi A, Lorito S (2007) Rupture process of the 2004 Sumatra-Andaman earthquake from tsunami waveform inversion. *Bull Seismol Soc Am* 97:S223–S231. <https://doi.org/10.1785/0120050627>
- Pietrzak J, Socquet A, Ham D, Simons W, Vigny C, Labeur RJ, Schrama E, Stelling G, Vatvani D (2007) Defining the source region of the Indian Ocean tsunami from GPS, altimeters, tide gauges and tsunami models. *Earth Planet Sci Lett* 261:49–64. <https://doi.org/10.1016/j.epsl.2007.06.002>
- Plafker G, Nishenko S, Cluff L, Syahrial M (2006) The cataclysmic 2004 tsunami on NW Sumatra – preliminary evidence for a near-field secondary source along the western Aceh basin. *Seismol Res Lett* 77:231
- Poisson B, Oliveros C, Pedreros R (2011) Is there a best source model of the Sumatra 2004 earthquake for simulating the consecutive tsunami? *Geophys J Int* 185:1365–1378. <https://doi.org/10.1111/j.1365-246X.2011.05009.x>
- Polet J, Kanamori H (2009) Tsunami earthquakes. In: Meyers A (ed) *Encyclopedia of complexity and systems science*. Springer, New York. https://doi.org/10.1007/978-0-387-30440-3_567
- Pritchard ME, Norabuena EO, Ji C, Boroschek R, Comte D, Simons M, Dixon TH, Rosen PA (2007) Geodetic, teleseismic, and strong motion constraints on slip from recent southern Peru subduction zone earthquakes. *J Geophys Res* 112:B03307. <https://doi.org/10.1029/2006JB004294>
- Rabinovich AB, Lobkovsky LI, Fine IV, Thomson RE, Ivelskaya TN, Kulikov EA (2008) Near-source observations and modeling of the Kuril Islands tsunamis of 15 November 2006 and 13 January 2007. *Adv Geosci* 14:105–116
- Rhie J, Dreger D, Burgmann R, Romanowicz B (2007) Slip of the 2004 Sumatra-Andaman earthquake from joint inversion of long-period global seismic waveforms and GPS static offsets. *Bull Seismol Soc Am* 97:S115–S127. <https://doi.org/10.1785/0120050620>
- Roeber V, Yamazaki Y, Cheung KF (2010) Resonance and impact of the 2009 Samoa tsunami around Tutuila, American Samoa. *Geophys Res Lett* 37:L21604. <https://doi.org/10.1029/2010GL044419>
- Romano F, Piatanesi A, Lorito S, D'Agostino N, Hirata K, Atzori S, Yamazaki Y, Cocco M (2012) Clues from joint inversion of tsunami and geodetic data of the 2011 Tohoku-Oki earthquake. *Sci Rep* 2:385. <https://doi.org/10.1038/srep00385>
- Romano F, Trasatti E, Lorito S, Piromallo C, Piatanesi A, Ito Y, Zhao D, Hirata K, Lanucara P, Cocco M (2014) Structural control on the Tohoku, earthquake rupture process investigated, by 3D FEM, tsunami and geodetic data. *Sci Rep* 5631:1–11. <https://doi.org/10.1038/srep05631>
- Romano F, Molinari I, Lorito S, Piatanesi A (2015) Source of the 6 February 2013 Mw 8.0 Santa Cruz Islands tsunami. *Nat Hazard Earth Syst Sci* 15:1371–1379. <https://doi.org/10.5194/nhess-15-1371-2015>
- Rothman D (1986) Automatic estimation of large residual statics corrections. *Geophysics* 51:332–346. <https://doi.org/10.1190/1.1442092>
- Satake K (2009) Tsunamis, inverse problem of. In: Meyers A (ed) *Encyclopedia of complexity and systems science*. Springer, New York
- Satake K (2010) Double trouble at Tonga. *Nature* 466:931–932. <https://doi.org/10.1038/466931a>
- Satake K, Fujii Y (2014) Review: source models of the 2011 Tohoku earthquake and long-term forecast of large earthquakes. *J Dis Res* 9:272–280
- Satake K, Fujii Y, Harada T, Namegaya Y (2013a) Time and space distribution of coseismic slip of the 2011 Tohoku earthquake as inferred from tsunami waveform

- data. *Bull Seismol Soc Am* 103:1473–1492. <https://doi.org/10.1785/0120120122>
- Satake K, Nishimura Y, Putra PS, Gusman AR, Sunendar H, Fujii Y, Tanioka Y, Latief H, Yulianto E (2013b) Tsunami source of the 2010 Mentawai, Indonesia earthquake inferred from tsunami field survey and waveform modeling. *Pure Appl Geophys* 170:1567–1582. <https://doi.org/10.1007/s00024-012-0536-y>
- Shearer P, Bürgmann R (2010) Lessons learned from the 2004 Sumatra-Andaman megathrust rupture. *Annu Rev Earth Planet Sci* 38:103–131. <https://doi.org/10.1146/annurev-earth-040809-152537>
- Sieh K, McKinnon EE, Pilarczyk JE, Chiang HW, Horton B, Rubin CM, Shen CC, Ismail N, Vane CH, Feener RM (2015) Penultimate predecessors of the 2004 Indian Ocean tsunami in Aceh, Sumatra: stratigraphic, archeological, and historical evidence. *J Geophys Res* 120:308–325. <https://doi.org/10.1002/2014JB011538>
- Singh SC, Carton H, Taponnier P, Hananto ND, Chauhan APS, Hartoyo D, Bayly M, Moeljopranoto S, Bunting T, Christie P, Lubis H, Martin J (2008) Seismic evidence for broken oceanic crust in the 2004 Sumatra earthquake epicentral region. *Nat Geosci* 1:777–781. <https://doi.org/10.1038/ngeo336>
- Sladen A, Tavera H, Simons M, Avouac JP, Konca AO, Perfettini H, Audin L, Fielding EJ, Ortega F, Cavagnoud R (2010) Source model of the 2007 Mw 8.0 Pisco, Peru earthquake – implications for seismogenic behavior of subduction megathrusts. *J Geophys Res* 115:B02405. <https://doi.org/10.1029/2009JB006429>
- Stein S, Okal EA (2011) The size of the 2011 Tohoku earthquake need not have been a surprise. *Eos* 92:227–228
- Stein S, Geller RJ, Liu M (2012) Why earthquake hazard maps often fail and what do about it. *Tectonophysics* 562–563:1–25. <https://doi.org/10.1016/j.tecto.2012.06.047>
- Strasser FO, Arango MC, Bommer JJ (2010) Scaling of the source dimensions of interface and intraslab subduction-zone earthquakes with moment magnitude. *Seismol Res Lett* 81:941–950
- Subarya C, Chlieh M, Prawirodirdjo L, Avouac JP, Bock Y, Sieh K, Meltzner AJ, Natawidjaja DH, McCaffrey R (2006) Plate-boundary deformation associated with the great Sumatra-Andaman earthquake. *Nature* 440:46–51. <https://doi.org/10.1038/nature04522>
- Synolakis C, Kanoglu U (2015) The Fukushima accident was preventable. *Phil Trans R Soc A* 373:2053
- Tanioka Y, Satake K (1996) Tsunami generation by horizontal displacement of ocean bottom. *Geophys Res Lett* 23:861–886
- Tappin DR, Grilli ST, Harris JC, Geller RJ, Masterlark T, Kirby JT, Shi F, Ma G, Thingbaijam KKS, Mai PM (2014) Did a submarine landslide contribute to the 2011 Tohoku tsunami? *Mar Geol* 357:344–361. <https://doi.org/10.1016/j.margeo.2014.09.043>
- Taylor FW, Briggs RW, Frohlich C, Brown A, Hornbach M, Papabatu AK, Meltzner AJ, Billy D (2008) Rupture across arc segment and plate boundaries in the 1 April 2007 Solomons earthquake. *Nat Geosci* 1:253–257. <https://doi.org/10.1038/ngeo159>
- Thomson RE, Rabinovich AB, Krassovski MV (2007) Double jeopardy: concurrent arrival of the 2004 Sumatra tsunami and storm-generated waves on the Atlantic coast of the United States and Canada. *Geophys Res Lett* 34:L15607. <https://doi.org/10.1029/2007GL030685>
- Titov V, Rabinovich AB, Mofjeld HO, Thomson RE, González FI (2005) The global reach of the 26 December 2004 Sumatra Tsunami. *Science* 309:2045–2048. <https://doi.org/10.1126/science/1114576>
- Tsai VC, Ampuero JP, Kanamori H, Stevenson DJ (2013) Estimating the effect of Earth elasticity and variable water density on tsunami speeds. *Geophys Res Lett* 40:492–496. <https://doi.org/10.1002/grl.50147>
- UNISDR (2015) Making development sustainable: the future of disaster risk management. Global assessment report on disaster risk reduction. United Nations Office for Disaster Risk Reduction (UNISDR), Geneva
- Vigny C, Simons WJF, Abu S, Bamphenyu R, Satirapod C et al (2005) Insight into the 2004 Sumatra-Andaman earthquake from GPS measurements in southeast Asia. *Nature* 436:201–206
- Wang K, Kinoshita M (2013) Dangers of being thin and weak. *Science* 342:2013. <https://doi.org/10.1126/science.1246518>
- Watada S, Kusumoto S, Satake K (2014) Traveltime delay and initial phase reversal of distant tsunamis coupled with the self-gravitating elastic Earth. *J Geophys Res* 119:4287–4310. <https://doi.org/10.1002/2013JB010841>
- Wei S, Helmberger D, Avouac JP (2013) Modeling the 2012 Wharton basin earthquakes off-Sumatra: complete lithospheric failure. *J Geophys Res* 118:3592–3609. <https://doi.org/10.1002/jgrb.50267>
- Yagi Y, Nakao A, Kasahara A (2012) Smooth and rapid slip near the Japan Trench during the 2011 Tohoku-Oki earthquake revealed by a hybrid back-projection method. *Earth Planet Sci Lett* 355–356:94–101. <https://doi.org/10.1016/j.epsl.2012.08.018>
- Yamazaki Y, Cheung KF (2011) Shelf resonance and impact of near-field tsunami generated by the 2010 Chile earthquake. *Geophys Res Lett* 38:L12605. <https://doi.org/10.1029/2011GL047508>
- Ye L, Lay T, Kanamori H (2012) The Sanriku-Oki low-seismicity region on the northern margin of the great 2011 Tohoku-Oki earthquake rupture. *J Geophys Res* 117:B02305. <https://doi.org/10.1029/2011JB008847>
- Yokota Y, Koketsu K, Fujii Y, Satake K, Sakai S, Shinohara M, Kanazawa T (2011) Joint inversion of strong motion, teleseismic, geodetic, and tsunami datasets for the rupture process of the 2011 Tohoku earthquake. *Geophys Res Lett* 38:L00G21. <https://doi.org/10.1029/2011GL050098>

- Yue H, Lay T (2011) Inversion of high-rate (1 sps) GPS data for rupture process of the 11 March 2011 Tohoku earthquake (Mw 9.1). *Geophys. Res. Lett.* 38:L00G09. <https://doi.org/10.1029/2011GL048700>
- Yue H, Lay T (2013) Source rupture models for the Mw 9.0 2011 Tohoku earthquake from joint inversions of high-rate geodetic and seismic data. *Bull Seismol Soc Am* 103:1242–1255. <https://doi.org/10.1785/0120120119>
- Yue H, Lay T, Koper KD (2012) En échelon and orthogonal fault ruptures of the 11 April 2012 great intraplate earthquakes. *Nature* 490:245–249. <https://doi.org/10.1038/nature11492>
- Yue H, Lay T, Rivera L, An C, Vigny C, Tong X, Báez Soto JC (2014) Localized fault slip to the trench in the 2010 Maule, Chile Mw = 8.8 earthquake from joint inversion of high-rate GPS, teleseismic body waves, InSAR, campaign GPS, and tsunami observations. *J Geophys Res* 119:1–19. <https://doi.org/10.1002/2014JB011340>
- Yue H, Lay T, Li L, Yamazaki Y, Cheung KF, Rivera L, Hill EM, Sieh K, Kongko W, Muhari A (2015) Validation of linearity assumptions for using tsunami waveforms in joint inversion of kinematic rupture models: application to the 2010 Mentawai Mw 7.8 tsunami earthquake. *J Geophys Res* 120:1728–1747. <https://doi.org/10.1002/2014JB011721>
- Zhao D, Huang Z, Umino N, Hasegawa A, Kanamori H (2011) Structural heterogeneity in the megathrust zone and mechanism of the 2011 Tohoku-oki earthquake (Mw 9.0). *Geophys Res Lett* 38:L17308. <https://doi.org/10.1029/2011GL048408>

Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings

Barry Hirshorn¹, Stuart Weinstein², Dailin Wang², Kanoa Koyanagi², Nathan Becker² and Charles McCreery²

¹NOAA/NWS, Ewa Beach, HI, USA

²NOAA/NWS/Pacific Tsunami Warning Center, Ewa Beach, HI, USA

Article Outline

Glossary
Definition of the Subject
Introduction
Tsunami Warning Center Operations
Seismic Methods
Earthquake Source Parameters
Future Directions
Bibliography

Glossary

CMT centroid moment tensor The CMT represents properties of the earthquake source derived from the seismic displacement of the Earth's crust that best reproduces the observed wave field generated by an earthquake and gives the average location in time and space of the earthquake energy release. The scalar seismic moment can be determined from the CMT.

Convolution A type of integral transform combining two signals to form a third signal or output. The output signal is typically viewed as a modified version of one of the two original signals, giving the area overlap between the two signals as a function of one of the original signals is translated with respect to the other. It is the single most important technique in digital signal processing. In the case of seismology, the two signals can be, e.g., the ground motion

as a function of time and the response of the seismometer, and the output is the seismogram.

Deconvolution Does the reverse of convolution. In the case of seismology, one uses deconvolution to remove the instrument response from the seismogram to recover the actual ground motion.

Deep earthquake An earthquake characterized by a hypocenter located more than 100 km below the Earth's surface.

Hypocenter The point within the Earth where the earthquake rupture starts. The epicenter is the projection of the hypocenter onto the Earth's surface.

Local tsunami warning A tsunami warning for distances less than 100 km from the source.

Magnitude

mB	The “broadband” body-wave magnitude, generally based on measurements of the amplitude of P-waves with periods in the 2 to 20 s range.
MS	The surface wave magnitude. MS is generally based on measurements of the amplitude of the surface (Love or Rayleigh) waves with periods of about 20 s. The US tsunami warning centers have applied a correction to the IASPEI formula that allows the estimation of MS closer to the epicenter at a period near 20 s.
ME	The “energy-magnitude” scale, derived from velocity power spectra.
Mm	The mantle wave magnitude, based on the measurement of the amplitude of surface waves with periods of 50–400 s.
MW	The moment magnitude, based on the estimation of the scalar seismic moment, M0.

Mwp	The moment magnitude based on the initial long-period P-waves.		definitions of the S-wave phases follow in the same manner.
ML	The local magnitude scale, based on the measurement of the maximum peak-to-peak amplitude observed on a Wood-Anderson seismogram, corrected for the decrease in amplitude with increasing epicentral distance. Generally based on the analysis of Sg, Lg, or Rg surface waves oscillating with periods observed out to 600 km from the earthquake's epicenter.	Seismic moment	The seismic moment, M_0 (expressed in units of force times distance, e.g., Newton-meters or dyne-cm), is the moment of either couple of an equivalent double couple point source representation of the slip across the fault area during the earthquake. Mathematically, the seismic moment, $M_0 = \mu A d$, where μ denotes the shear rigidity or resistance of the faulting material to shearing forces, A represents the area of the fault plane over which the slip occurs, and d represents the average coseismic slip across A.
pMag	A magnitude scale based on the average of the absolute values of the first three half cycles of the P-waves recorded at local distances.		Elastic waves generated by movements of the earth's crust that propagate as radiated seismic energy, ER.
Marigram	A recording of sea-level variations obtained by tide gauges.	Seismic waves	Waves that propagate along the surface boundary of a medium, e.g., along the surface of the earth.
Regional tsunami warning	A tsunami warning for distances up to 1000 km from the source.	Seismic surface waves	An earthquake characterized by a hypocenter located within 100 km of the Earth's surface.
Seismic body waves	Waves that propagate through the interior of an unbounded continuum. Primary waves (P-waves) are longitudinal body waves that shake the ground in a direction parallel with the direction of travel. Secondary body waves (S-waves) are shear waves that shake the ground in a direction perpendicular to the direction of travel. There are other types of arrivals (also known as phases) visible on seismographs corresponding to reflections of P- and S-waves from the earth's surface: The pP phase is a P-wave that travels upward from the hypocenter and reflects once off the surface, and the PP phase is a P-wave that travels downward from the epicenter and reflects once off of the surface. The	Shallow earthquake	A tsunami warning for distances greater than 1000 km away from the source.
		Teletsunami warning	A series of water waves generated by any rapid, large-scale disturbance of the sea. Most tsunamis are generated by seafloor displacements from large undersea earthquakes, but they can also be caused by large submarine landslides, volcanic eruptions, calving of glaciers, and even by meteorite impacts into the ocean.
		Tsunami	An earthquake that generates a much larger tsunami than expected given its magnitude. A tsunami warning system consists of a tsunami warning
		Tsunami earthquake	

Tsunami warning system	center such as the Pacific Tsunami Warning Center (PTWC), a formal response structure that includes civil defense authorities and government officials, and an education program that brings a minimum level of awareness and education to the coastal populations at risk.
W-phase	A distinct long-period, up to 1000 s, phase which arrives before the S phase. Kanamori (1993) first observed it on the displacement records of the 1992 Nicaragua tsunami earthquake. The W-phase is so-called because it propagates in a manner similar to that of the acoustic whispering gallery mode (Kanamori 1993).

approximately 7 min, whereas a tsunami takes about 5½ h to travel the same distance. Although over 200 sea-level stations, reporting in near real time, are operating in the Pacific, it may take an hour or more, depending on the location of the epicenter, before the existence (or not) of an actual tsunami is confirmed. In other ocean basins where the density of sea-level instruments reporting data in near real time is less, the delay in tsunami detection is correspondingly longer. In addition, global, regional, and local seismic networks, and the infrastructure needed to process the large amounts of seismic data that they record, are widespread. For these reasons, tsunami warning centers provide their initial tsunami warnings to coastal populations based entirely on seismic data.

Introduction

Definition of the Subject

Tsunamis are among nature's most destructive natural hazards. Typically generated by large, under-water earthquakes near the Earth's surface, tsunamis can cross an ocean basin in a matter of hours. Although difficult to detect, and not dangerous while propagating in deep water, tsunamis can unleash immense destructive power when they reach coastal areas. With advance warning, coastal populations can be alerted to move to higher ground and away from the coast, saving many lives. Unfortunately, due to the lack of a tsunami warning system in the Indian Ocean, the Sumatra earthquake and tsunami of Dec. 26, 2004, killed over 250,000 people, with thousands of lives lost as far away as East Africa many hours after the earthquake occurred. Had a tsunami warning system been in place, many lives could have been saved.

As fast as tsunami waves are, seismic waves travel at speeds more than 40 times greater. Because of this large difference in speed, scientists rely on seismic methods to detect the possibility of tsunami generation and to warn coastal populations of an approaching tsunami well in advance of its arrival. Seismic P-waves, for example, travel from Alaska to Hawaii in

Any mechanism that causes a sudden displacement of the ocean's surface affecting a significant volume of water can produce a tsunami. Undersea earthquakes, landslides, volcanic explosions, calving of icebergs, and even meteorite impacts can generate tsunamis. However, the majority of tsunamis are generated by earthquakes. Not uncommon are earthquakes that trigger landslides, so that both the displacement of the crust due to the earthquake and the landslide each contribute to the generation and size of the tsunami. Tsunamis are a devastating natural, high fatality hazard (Bryant 2001). In the absence of a proper tsunami warning system, a destructive tsunami will cause death and destruction as it encounters coastal areas while propagating across an entire ocean basin as it did in the case of the Sumatra tsunami of December 2004.

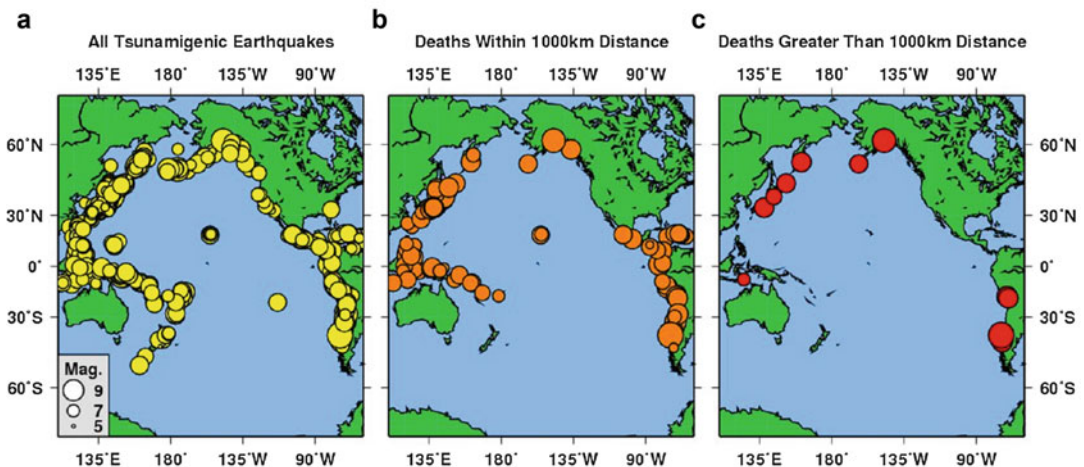
Although tsunamis propagate in deep water with speeds exceeding 900 km/h, they are hard to detect in the open ocean. For example, the first wave of the great 2004 Sumatra tsunami had a wave height of only 1 m in deep water (depth > 500 m) (Gower 2005) and a wavelength on the order of several hundred kilometers. Consequently, people aboard ocean vessels did not feel the accelerations caused by the tsunami as it passed under them. However, as the speed, v (in m/s), of a

tsunami is governed by the simple relation $v = \sqrt{gh}$, where g is the acceleration of gravity (in m/s^2) and h is the thickness of the water column (in m), the tsunami will slow down as it propagates into shallow water. At this point, the wave speed and wavelength decrease, causing the wave height to increase. Depending on the nature of the tsunami, and the shape and bathymetry of the coastal area, the tsunami wave height can be greatly amplified, magnifying its destructive power.

Because earthquakes generate most tsunamis, and seismic waves travel more than 40 times faster than tsunamis, the first indication that a tsunami may have been generated is the earthquake itself. Depending on the earthquake's location (undersea or inland), depth (shallow or deep) in the Earth's crust, and on its magnitude, a tsunami warning center may issue an official message product. If the earthquake is a shallow, undersea earthquake, the severity indicated by the initial message will depend upon the magnitude of the earthquake. The more rapidly and accurately the tsunami warning center can characterize the earthquake source, the faster it can make the initial evaluation of the tsunamigenic potential of the earthquake.

While some tsunamis are destructive, most are rather small, producing few if any casualties and little or no damage, although they are easily observable on marigrams (Fig. 1). On the basis of how widespread their effects are, tsunamis can be classified as local (within 100 km of the epicenter), regional (up to 1000 km from the epicenter), or teletsunamis (greater than 1000 km from the epicenter). Warning centers designed to respond to tsunami threats on each of these scales now exist in every major ocean basin.

The Pacific Tsunami Warning Center (PTWC) provides basin-wide warnings to the coastal areas of the Pacific and Caribbean basins. The PTWC also functions as a local tsunami warning center for Hawaii, Puerto Rico, the Virgin Islands, Guam, and American Samoa. Other local tsunami warning centers include the CPPT (French Polynesia Tsunami Warning Center), which is based in Tahiti, GFZ-Indonesia, which is based in Jakarta which provides local warnings for Indonesia, and Japan's JMA (Japan Meteorological Agency) which provides local tsunami warnings to Japan. Examples of regional warning centers include the Japan Meteorological Agency (JMA), which provides regional tsunami warnings to the



Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings, Fig. 1 Epicenters of tsunamigenic earthquakes occurring in the Pacific since 1 A.D. Of those earthquakes that do produce a tsunami (a), most tsunamis cause no damage. Most events that cause casualties and/or damage do so within 1000 km of the

epicenter (b), leaving only a few great earthquake sources that generated tsunamis which caused casualties and/or damage more than a 1000 km from the epicenter. (c) Data provided by the NOAA National Geophysical Data Center (NGDC) (www.ngdc.noaa.gov/hazard/tsu.html)

Northwest Pacific, and the National Tsunami Warning Center (NTWC), which provides regional warnings to the US mainland coasts and the west coast of Canada.

The tsunami warning centers themselves are not complete tsunami warning systems; they are simply the first line of defense within these warning systems. A tsunami warning system consists of three main components: (a) the tsunami warning center, (b) the emergency management/civil defense authorities who receive tsunami warning center message products and mobilize the public, and (c) the coastal populations themselves, who must be educated on how to respond to tsunami emergencies. If any of these three components are lacking, the tsunami warning system will fail. Unfortunately, none of these components existed in the Indian Ocean at the time of the December 2004 Sumatra earthquake and tsunami. However, in the decade since this disaster, Indonesia, Australia, and India have created tsunami warning systems in the Indian Ocean basin.

The greatest challenge for a tsunami warning system, particularly in the near field, is rapid identification of slow (in terms of fault rupture speed) “tsunami” earthquakes. Tsunami earthquakes are so-called because they generate much larger tsunamis than expected from their magnitude (Kanamori 1972). In a well-functioning tsunami warning system, residents in coastal areas are educated to immediately move inland, and onto higher ground if they feel strong ground shaking, and not wait for an official tsunami alert (Fryer et al. 2005). However, because a tsunami earthquake produces much less radiated high-frequency energy than normal, even a large (in terms of moment magnitude) tsunami earthquake may not be strongly felt in the near field, so that this strategy of having people self-evacuate upon feeling strong ground shaking will not work. This was, unfortunately, made clear by the Java “Tsunami” earthquake of July 17, 2006. The tsunami generated by this earthquake killed at least 500 people, as many residents in coastal areas near the earthquake did not feel strong shaking (Widjo et al. 2006). Tsunami warning centers need to be able to properly detect the occurrence of these tsunami earthquakes as soon as possible after they begin.

Tsunami Warning Center Operations

Tsunami warning centers function much like seismic observatories in detecting, locating, and characterizing earthquakes, with the difference that they are primarily interested in very large earthquakes and that their earthquake parameterization needs to be very fast, sometimes sacrificing precision. Depending on the earthquake’s location (underwater vs. inland), depth below the surface, and magnitude, tsunami warning centers may issue an official message product to advise emergency management authorities within the warning center’s AOR (area of responsibility) of the occurrence of a large earthquake and its potential for generating a tsunami. The PTWC, located in Honolulu, Hawaii, provides such notice regarding the potential or confirmed generation of a destructive tsunami for most of the Pacific Ocean and Caribbean Sea and their adjacent seas. The PTWC also provided tsunami threat information for the Indian Ocean basin following the December 26, 2004, Indian Ocean tsunami until the end of April 2013.

PTWC’s international message products and criteria were substantially changed in 2014, after several years of consultation with and training of the Pacific Tsunami Warning System (PTWS) Member States. The new products were created in an effort to reduce the amount of over-warning and to give more authority to the disaster management organizations of the Member States to formulate what specific alerts and actions to take along their coastlines. Among the important changes are the adoption of new terminology that provides threat levels instead of alert levels like “warning” or “watch” and the inclusion of tsunami wave amplitude forecasts in the message products. Under the new guidelines, the first bulletin issued assesses the potential of the earthquake to generate a destructive tsunami based solely on the earthquake location, depth, and magnitude, just as the previous products did. However, after PTWC obtains a reliable CMT, about 25 minutes post origin time, subsequent products base tsunami threat levels on forecasted and observed wave heights. The following provides a general description of PTWC’s products for the Pacific.

Observatory Message: The PTWC sends an observatory message to certain seismological observatories and organizations for any earthquake with a magnitude of about 5.8 or greater. In areas with dense seismic networks, the PTWC may issue an observatory message for earthquakes as small as magnitude 4. This unofficial message contains only the earthquake's preliminary epicentral location, origin time, depth, magnitude, and a list of stations used in computing these parameters. These messages contain no evaluation regarding the seismic or tsunami hazard.

Tsunami Information Statement: The PTWC issues this message product for any earthquake in the vicinity of the Pacific basin with a magnitude in the range $6.5 \leq M_w \leq 7.0$ or for larger earthquakes when they are located too far inland or too deep inside the earth to pose a tsunami threat.

Tsunami Threat Message: The PTWC issues this message product for any shallow (depth < 100 km) undersea or nearshore earthquake in or near the vicinity of the Pacific basin with a magnitude $M_w \geq 7.1$. If the magnitude is in the range $7.1 \leq M_w \leq 7.5$, the tsunami threat message will indicate a possible tsunami threat to coasts located within 300 km of the epicenter. If the magnitude is in the range $7.6 \leq M_w \leq 7.8$, the tsunami threat message will indicate a possible tsunami threat to coasts located within 1000 km of the epicenter, and finally, if the magnitude is $M_w \geq 7.9$, the tsunami threat message will indicate a possible tsunami threat to coasts located within 3 h tsunami travel time of the earthquake's epicenter.

After a reliable CMT is obtained, subsequent tsunami threat messages will place coastlines in "threat levels" based on the maximum forecast wave heights generated by RIFT (see section on "Real-Time Tsunami Forecasting") and any sea-level observations that are available. The threat levels assigned to a coastline are described below:

1. No threat: Maximum forecast wave amplitude is below 30 cm.
2. Maximum forecast wave amplitude (H) is in the range $30 \text{ cm} \leq H < 1 \text{ m}$.

3. Maximum forecast wave amplitude (H) is in the range $1 \text{ m} < H \leq 3 \text{ m}$.
4. Maximum forecast wave amplitude greater than 3 m.

In general, threat level 2 indicates the possibility of dangerous offshore currents as well as some flooding of beaches and harbors. Threat levels 3 and 4 generally indicate an inundation hazard, with level 4 indicating the possibility of extensive inundation along the coast by the tsunami.

Supplemental tsunami threat messages are issued with updated sea-level observations and other information until the tsunami hazard has generally passed at which time a final tsunami threat message is issued.

In consultation with the Caribbean tsunami warning system (CARIBE-EWS), the PTWC also issues the following products for the Caribbean Ocean basin:

Tsunami Information Statement: The PTWC issues this message product for any earthquake in the vicinity of the Caribbean Sea, with a magnitude in the range $6.0 \leq M_w \leq 7.0$, or in the Atlantic with a magnitude in the range $6.5 \leq M_w \leq 7.8$ or for larger Caribbean or Atlantic earthquakes when they are located too far inland or too deep inside the earth to pose a tsunami threat.

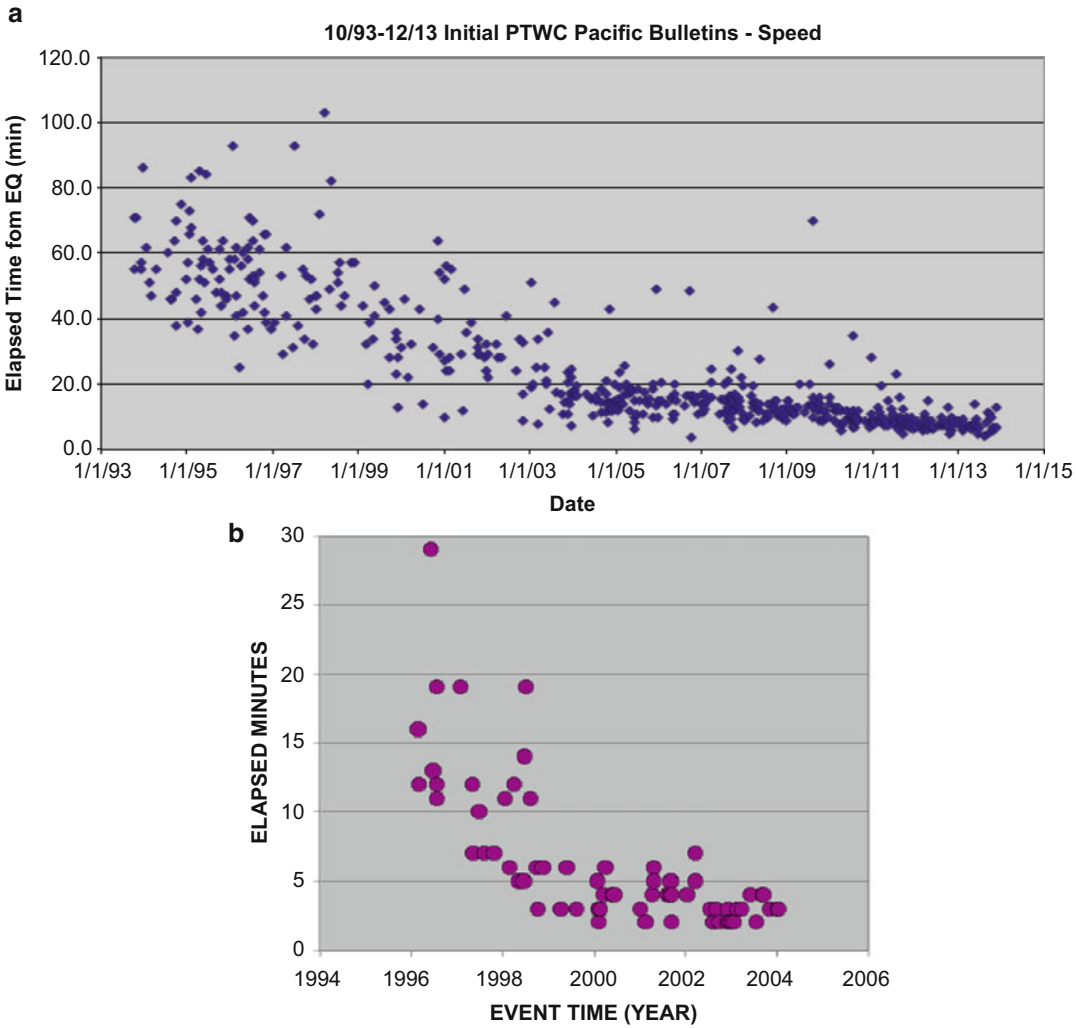
Tsunami Threat Message: Tsunami threat messages issued for the Caribbean Ocean basin or Atlantic Ocean basin are similar, and the magnitude criteria and threat levels are the same as in the above paragraph describing the Pacific Ocean basin tsunami threat messages.

Coastlines close to the earthquake epicenter can experience tsunami waves within 2 to 30 min after the earthquake; hence a local tsunami warning needs to be issued within a few minutes to be effective. This requires access to real-time data provided by a dense local network of seismic stations near the epicenter to allow both the rapid location and source characterization of the earthquake. In the case of the Hawaii region, the PTWC uses data from its own local seismic network and from the dense seismic network

maintained by the USGS Hawaii Volcano Observatory (HVO) to rapidly detect and characterize local Hawaii earthquakes. These data, combined with automatic local Earthquake detection, association, and paging based on initial p-wave magnitude estimations, introduced at the PTWC in 1999–2000 (Allen 1978, 1982; Johnson et al. 1994, 1995; Hirshorn et al. 1993) enabled the PTWC to issue an information bulletin to the state of Hawaii and to the Pacific basin, for the October 15, 2006, Kiholo Bay earthquake

(M_w 6.7) within 3 min of the origin time of the earthquake (Hirshorn 2007) (Fig. 2b). However, the PTWC cannot issue timely local Tsunami warnings for populations in the immediate vicinity of large earthquakes outside of Hawaii, Puerto Rico, or the Virgin Islands because PTWC lacks access to dense enough local seismic networks outside of these regions (Fig. 2a).

Japan maintains an extremely dense seismic network and can therefore rapidly issue warnings for offshore earthquakes. Spurred on by the



Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings, Fig. 2 Elapsed time from earthquake origin time to time of issuance of the first official message product for (a) teleseisms (earthquakes

that occur outside of the Hawaiian Islands region) and (b) for earthquakes that occur within the Hawaiian Islands. The PTWC’s mean time after origin for Pacific Bulletins has been 5–6 minutes from 2017 through 2018

December 2004 Sumatra and Chile 2010 earthquakes, several other nations such as Indonesia, Australia, Chile, and New Zealand, for example, have all rapidly developed their seismic networks in an effort to improve their local tsunami warning capabilities for earthquakes along their shores.

Seismic Methods

To rapidly detect, locate, and characterize the source of earthquakes occurring around the world, tsunami warning centers rely on the Global Seismic Network (GSN USGS/IRIS), the USGS/NEIC, and a number of other seismic data contributors worldwide. It is this unfettered access to real-time seismic data supplied by a number of different networks that makes basin-wide tsunami warning centers possible. To rapidly deal with the threat posed by locally generated tsunamis to the state of Hawaii, PTWC processes seismic data from about 80 stations located throughout the Hawaiian Islands. The USGS HVO's dense seismic network supplies most of this data. The US tsunami warning centers use the Earthworm software developed by the USGS to import and export seismic data (Johnson et al. 1995).

PTWC duty scientists receive automatic pages within a few minutes of the beginning of any earthquake with a M_W of approximately 5.5 or greater. The system generating these pages uses Evan's and Allen's (1983) teleseismic event detection algorithm, adapted for broadband data by Wither's (1998) and Whitmore's and Blackford's (2002) teleseismic picker (which picks the first p-wave arrival time) and phase associator (which associates those arrival times belonging to a single earthquake together). In the Hawaiian Islands, Puerto Rico, and the Virgin Islands, Earthworm (Johnson et al. 1995) automatically notifies duty scientists of earthquakes with magnitudes larger than 3.5 within 10 to 20 s of the earthquake's origin time, locates the hypocenter, and provides a first estimate of the earthquake's magnitude and other source parameters in real time (Allen 1978, 1982; Johnson et al. 1994, 1995). PTWC duty scientists then refine and supplement the automated real-time hypocenter location and/or magnitude estimates.

Determining the earthquake's depth is particularly important as earthquakes occurring at depths greater than 100 km generally do not cause dangerous tsunamis.

Earthquakes are located using P-wave arrival times recorded at a number of seismic instruments. As both the locations of seismic instruments and P-wave travel times as a function of distance are well known, a process analogous to triangulation is used to locate the earthquake. The pickers and associators perform these functions automatically on a continuous basis.

While the depth of the earthquake can be estimated on the basis of P-times alone, a more robust result often requires the addition of depth phases such as pP, which is a P-wave that travels directly up to the earth's surface from the earthquake source and reflects once off of that surface before arriving at the seismometer. The duty scientists use pP arrival times to refine hypocentral depths of distant earthquakes (teleseisms). For earthquakes in Hawaii, Puerto Rico, and the Virgin Islands, observed at local distances, the S-wave arrival time would be useful for constraining earthquake depth.

Seismologists use a panoply of different magnitude scales to characterize the seismic source. These different methods examine different parts of the seismic wave train, such as short- and long-period body waves (seismic waves that travel through the earth's interior like the P- and S-waves) and longer-period surface waves (slower seismic waves that are constrained to travel along the earth's surface). Most of these magnitude scales were developed to estimate the energy released by the earthquake as radiated seismic wave energy, E_R . Traditional magnitude measures such as M_L (Richter 1935) and m_b (a shorter-period variant of m_B Gutenberg (1945c; Gutenberg and Richter 1956a)) examine high-frequency body waves. Gutenberg's surface wave magnitude M_S (Gutenberg 1945a; Vanek et al. 1962) is derived from longer-period surface waves. A relatively new and quick method, M_{wp} , analyzes long-period P-waves (Tsuboi et al. 1995, 1999; Whitmore et al. 2002). The M_{wp} magnitude now provides the basis used to decide which, if any, official message product to issue, replacing the M_S method that the PTWC had been using for over

50 years. For large earthquakes, duty scientists also routinely estimate the mantle magnitude M_m , a very long-period surface wave magnitude based on mantle waves with periods in the range 50–410 s (Okal and Talandier 1989). The relationship between these magnitudes, each looking at different parts of the seismic wave spectrum of an earthquake, can be used to characterize the earthquake source (Aki 1967; Brune 1970, 1971; Kanamori 1977, 1978, 1983; Kanamori and Kikuchi 1993).

Today, the PTWC's best estimate of M_W comes from the inversion of the W-phase to obtain the CMT (Kanamori and Rivera 2008). The W-phase CMT calculation is triggered by any PTWC preliminary earthquake message for an earthquake with magnitude >6.8 . A W-phase-based CMT and M_W are then available within ~ 25 min after the earthquake origin time. The PTWC is testing a regional WCMT which will cut the time required to obtain a robust CMT down to 15 minutes. The PTWC expects to operationalize the regional WCMT in early 2020.

When evaluating the tsunamigenic potential of an earthquake, PTWC duty scientists compute the quantity $\log_{10}(E_R/M_0)$, known as “theta” Θ , where E_R is the radiated energy carried by the high-frequency p-waves and M_0 is the seismic moment (Aki 1966). Newman and Okal (1998) showed that Θ is anomalously small for tsunami earthquakes.

Since the mid-1990s, both US tsunami warning centers response times to potentially tsunamigenic teleseisms have decreased dramatically due to the much larger amounts of seismic data that they now receive and to the switch from the slower M_S magnitude method to the faster and more accurate M_{wp} moment magnitude method as the basis for issuing messages (Fig. 2a). Improved data analysis methods and increased amounts of local seismic data have also significantly decreased the PTWC's response time to earthquakes occurring in the Hawaiian Islands region (Fig. 2b).

Earthquake Source Parameters

A fundamental problem with traditional magnitude estimates, such as M_L , m_b , and M_S , is that they are based on the amplitudes of relatively short-period seismic waves, with periods usually

less than 3 s for m_b and M_L and 20 s for M_S . When the largest rupture dimension of the earthquake exceeds the wavelength of these seismic waves, which, for example, is about ~ 50 km for the 20 s period surface waves used for M_S (Kanamori 1977, 1978, 1983), these magnitude values will start to “saturate.” Saturation in this case means that these magnitude methods will underestimate M_W when the periods of the waves on which they are based are shorter than the corner period of the earthquake's seismic wave spectrum (Aki 1967; Brune 1970, 1971) (see Fig. 3).

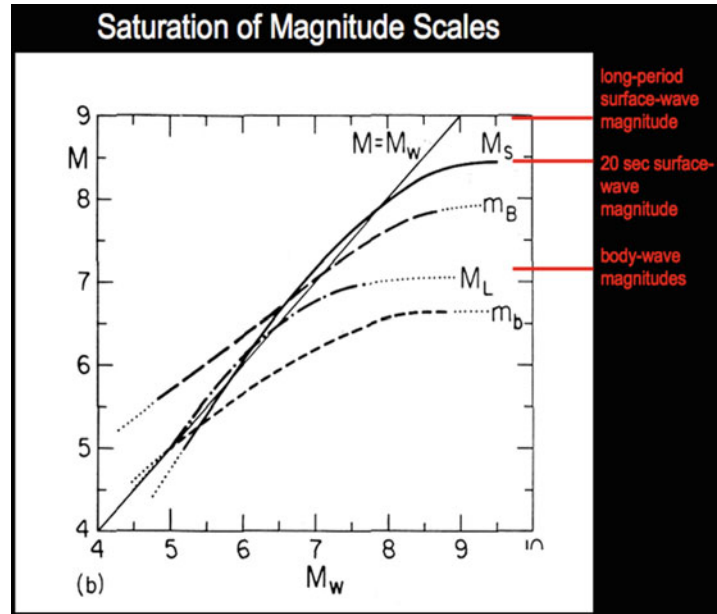
Another equivalent explanation is that these magnitude methods, which look at waves with periods of a fraction of a second to a few tens of seconds, cannot sample enough of the energy released by an earthquake whose source duration (the length of time over which the rupture occurs) is many times longer than the periods used by these methods. As the earthquake becomes very large, one must examine longer-period waves to avoid saturation.

K. Aki used a spectral representation to establish that earthquakes of varying size had similarly shaped spectra, differing primarily in their low-frequency amplitudes, proportional to seismic moment, and in the location of a given spectra's “characteristic frequency” (corner frequency of the source spectrum) (Aki 1967). He related this corner frequency to the characteristic length scale of the earthquake's rupture area (Aki 1967). Brune (1970, 1971) and Savage (1972) also related the corner frequency to the dimensions of the fault plane.

To circumvent the saturation problem, Kanamori defined the moment magnitude M_W (Kanamori 1977), in terms of a minimum estimate of the total coseismic strain energy drop, W_0 , via Gutenberg and Richter's energy-magnitude relationship (Gutenberg and Richter 1956b). M_W measures the work required to rupture the fault, as computed from the seismic moment, M_0 , assuming (1) that the coseismic stress drops associated with large earthquakes are approximately constant and (2) that the stress release during an earthquake is about the same as the kinetic frictional stress during faulting. Hanks and Kanamori (1979) discuss M_W and its agreement with the M_L and M_S magnitude scales in their unsaturated ranges,

Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings,

Fig. 3 Saturation of different classical magnitude scales with respect to non-saturating moment magnitude according to Kanamori. Note that m_B refers to the original Gutenberg-Richter (1945b, c) body-wave magnitude scale based on amplitude measurements made on medium-period broadband instruments. It saturates at larger magnitudes when compared to the short-period-based m_b



while Kanamori (1983) discussed the average relationship of M_w with m_b and m_B , also in the range where these magnitudes saturate (see Fig. 3).

Traditional Amplitude-Based Magnitudes at the PTWC

Hirshorn et al. (1993) developed a short-period P-wave magnitude scale, called *pMag*, based on the average of the absolute values of the amplitudes of the first three half cycles of the initial p-waves recorded, at local distances, on short-period seismometers. The *pMag* scale is based on the assumption that locally recorded initial P-wave amplitudes share a common decay curve shape in a given geographic area, independent of the magnitude of the earthquake, as Richter showed was the case for S-waves in his derivation of the M_L magnitude scale for Southern California (Richter 1935). Lindh and Hirshorn incorporated *pMag* into Carl Johnson's (1994) local p-wave associator, enabling automatic pages that contain the hypocentral parameters and the lower bound magnitude estimate provided by *pMag*, within 10 to 20 s of an earthquake's origin time.

PTWC now uses local algorithms from the stock Earthworm system (Johnson et al. 1995) to provide earthquake hypocenters and local

synthetic Wood-Anderson magnitudes, M_L , for earthquakes occurring in Hawaii and the Puerto Rico/Virgin Islands Region. Depending on the density of the seismic network near the epicenter, this information is available from 20 to 60 s after earthquake origin time.

Due to the sparseness of the Caribbean network, locations from our local system are generally augmented by manual picks and stations further from the epicenter to provide additional azimuthal control for larger events. The local systems generate fairly reliable M_L magnitudes up to approximately M_6 , above which M_{wp} (Tsuboi et al. 1995, 1999; Whitmore and Sokolowski 2002) or W-phase (Kanamori and Rivera 2008)-based estimates of M_w are necessary for accurate source size estimation.

PTWC's body-wave magnitude method, *bMag*, has similarities to the intermediate period broadband body-wave m_B magnitude, as defined by the IASPEI. The IASPEI m_B (Bormann et al. 2002; Bormann and Saul 2008b) is based on Gutenberg's (1945b, c) and Gutenberg and Richter's m_B (Gutenberg and Richter 1956a, b). *bMag* uses a 90 s window of the broadband vertical component seismogram starting 30 s prior to the arrival of the P-wave. This window is band-passed filtered between 0.3 and 5 s. The largest amplitude and its

period found in the 60 s after the first P-wave arrival are used in the magnitude formula. In PTWC's implementation, the formula used is the same as Gutenberg and Richter's (Gutenberg and Richter 1956a, b) relation, adopted by IASPEI for m_B :

$$bMag = \log_{10}(A_{\max}/T_{\max}) + Q(\Delta, z)$$

where Δ is the epicentral distance ($15 \leq \Delta \leq 90^\circ$), z is the hypocentral depth (in km), $A_{\max}$ is the maximum wave amplitude obtained from the band-pass filtered record, and $T_{\max}$ is the period of the wave with that maximum amplitude. Gutenberg and Richter's (Gutenberg and Richter 1956b) table of $Q(\Delta, z)$ is used to provide the distance and depth corrections. The largest amplitude found in the 30 s prior to the P-wave arrival time is used as the basis for the signal-to-noise ratio. $bMag$ differs from the IASPEI m_B (Bormann et al. 2006; Bormann and Saul 2008a) in three respects:

1. $bMag$ uses the largest amplitude wave in the first 30 sec after the initial P-wave arrival time.
2. $bMag$ uses a slightly different distance range ($15 \leq \Delta \leq 90^\circ$).
3. For $bMag$, the seismogram is band-pass filtered using the band .3 to 5 s.

$bMag$ will saturate at lower magnitudes than M_S does, so it is of limited use for large earthquakes. However, $bMag$ is still useful for three main reasons:

1. Unlike M_S , $bMag$ has a correction for the depth of the event's hypocenter.
2. $bMag$ is useful for determining the magnitude of moderate earthquakes that occur after much larger earthquakes, e.g., when longer-period energy is still present in the signal from the earlier, larger event that can adversely affect magnitude methods based on those longer periods.
3. By comparison with magnitudes based on longer periods, such as M_H , the shorter-period-based $bMag$ can also provide a way to detect slow or tsunami earthquakes.

To compute M_S (as first proposed by Gutenberg 1945a and later revised by Vanek et al. 1962) at the PTWC, we first band-pass filter 14 min of the broadband velocity seismogram, from 16 to 23 s, starting 3 min before the expected arrival time of the surface waves. We then apply the following equation, similar to the IASPEI (Bormann et al. 2002; Vanek et al. 1962) formula:

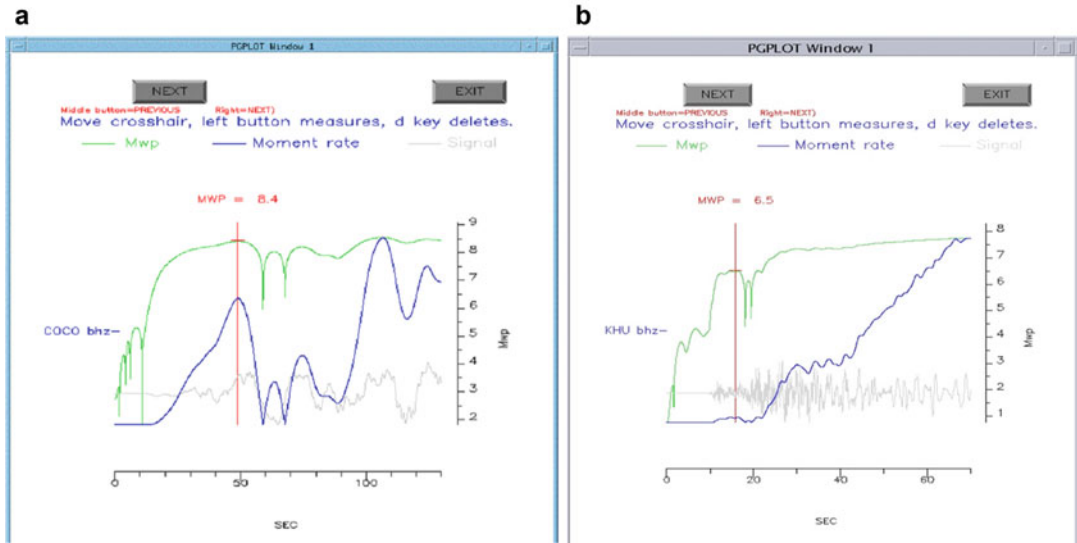
$$M_S = \log_{10}(A_{\max}/T_{\max}) + 1.66 \log_{10}(\Delta) + 3.3 \\ + \text{correction}$$

The correction term is 0 for epicentral distances, Δ , greater than 16° , and $0.53-0.033\Delta$ for Δ less than 16° . This correction term allows the US TWCs to compute the M_S magnitudes from stations as close as 600 km to the epicenter, at a period of 20 s. Note that in the IASPEI implementation of M_S , because it considers a much greater period range, from 3 to 60 s for $T_{\max}$, there is no need for our correction term. Both the PTWC's and the IASPEI's M_S implementations are susceptible to saturation effects as the magnitude reaches the high 7's.

Although the TWCs no longer use M_S as the basis for issuing bulletins, it is still helpful in discriminating deep from shallow earthquakes and for comparing the amount of 20 s radiated energy to the amounts of radiated energy at other periods. Deep earthquakes do not excite large surface waves. Hence if $bMag > M_S$, then the hypocenter is likely to be deep.

The Mwp Method

The broadband P-wave moment magnitude, M_{wp} , has replaced M_S as the magnitude upon which both US TWC's initial tsunami messages are based (Tsuboi et al. 1995, 1999; Whitmore et al. 2002). This is because M_{wp} is obtained much earlier than M_S , which is based on the slower traveling surface waves, and because M_{wp} examines much longer-period waves than the 20 s surface waves used by M_S , making it less susceptible to the saturation effects discussed above. M_{wp} , as implemented at the PTWC, examines up to the first 120 s of the vertical component, broadband velocity



Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings, Fig. 4 (a) Is the first 2 min of the broadband vertical velocity seismogram beginning with the p-wave arrival (gray) recorded by the GSN USGS/IRIS broadband station COCO, on Cocos Island, ~15 degrees south of the epicenter of the M_w 9.2 Sumatra earthquake of Dec. 26, 2004. Note that this portion of the broadband velocity seismogram is not clipped.

This instrument, a KS54000, has a flat frequency response to velocity to a period of about 350 s. The blue trace is the integrated displacement record (doubly integrated velocity) $\sim Mo(t)$, and the green trace, $M_{wp}(t)$, is M_{wp} as a function of time. (b) Shows these results at KHU, 77 km from the epicenter of the M_w 6.7 (GCMT) Kiholo Bay Hawaii earthquake of October 16, 2006

seismogram, beginning at the P-wave arrival time (gray traces in Fig. 4).

The derivation of M_{wp} assumes that we can estimate the seismic moment, M_0 , from the initial portion of the far-field P-wave observed on the vertical broadband displacement waveform, $u_Z(x_r, t)$, using

$$M_0 = (4\pi\rho\alpha^3 r / F^P) \text{Max} \left| \int u_Z(x_r, t) dt \right|,$$

where ρ and α are, respectively, the density and P-wave velocity averaged along the propagation path, r is the epicentral distance, and F^P is the earthquake source radiation pattern (Tsuboi et al. 1995, 1999). At the PTWC, before early 2011, we followed Tsuboi et al. (1995), by approximating $\text{Max}|\int u_Z(x_r, t) dt|$ by the first significant or “big” peak in the absolute value of the integrated displacement record. It is best to use velocity, $v(t)$, seismograms from broadband seismometers with flat responses out to at least 300–350 s, as there is then no need to deconvolve the instrument response from the data. Instead, we simply scale

the data by a gain factor in the time domain, because we can assume that the amplitude of the instrument response function is flat over the frequency band of interest. For data from very long period broadband seismometers, such as the STS-1, Trillium 360, etc. which have flat spectral responses to velocity up to at least 300 s period, M_{wp} closely approximates M_w for all but the very largest and/or slowest earthquakes.

Beginning with the raw velocity seismogram, $v(t)$, we subtract the average of 300 s of $v(t)$, ending 60 s before the initial P-wave arrival time, from the entire seismogram. We then integrate twice and multiply the absolute value of each data point by $4\pi\rho\alpha^3 r$ to obtain an approximation to $M_0(t)$ in N-m (the blue traces in Fig. 4). We then apply H. Kanamori’s moment magnitude formula (Kanamori 1977):

$$M_w(t) = (\log_{10} M_0(t) - 9.1)/1.5$$

to $M_0(t)$, to calculate $M_{wp}(t)$. To correct for the radiation pattern, F^P , we then add 0.2 to the average of these individual M_{wp} values,

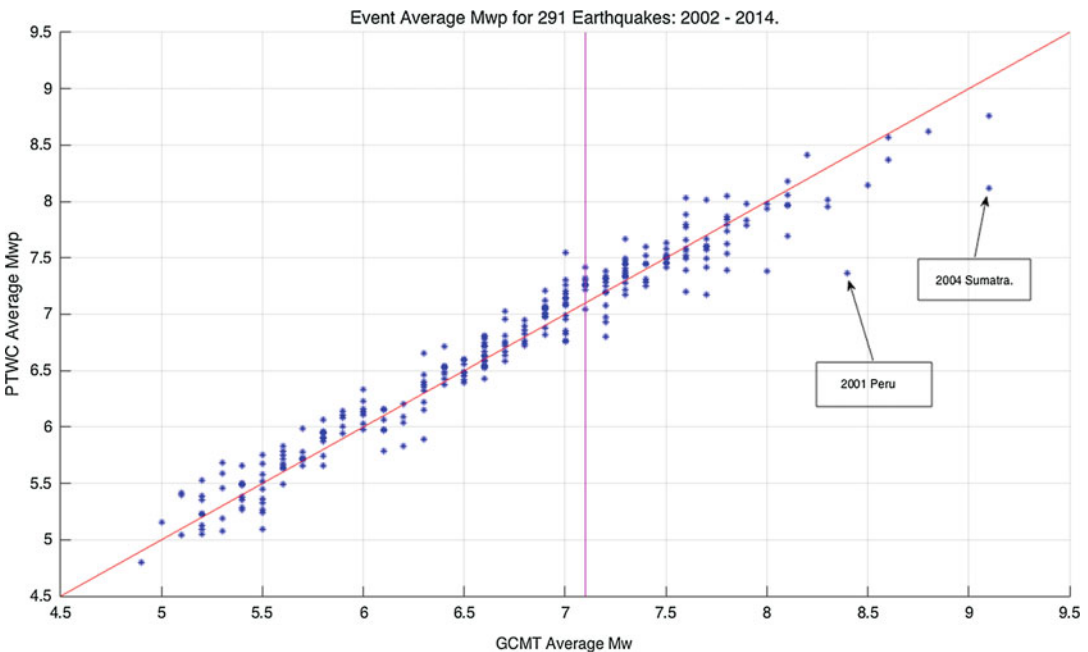
each obtained at a different azimuth and distance from the epicenter. This is because $\int (F^P)^2 d\Omega = 4/15$ where Ω is the azimuthal angle of the observation around the epicenter; thus $F^P \propto \sqrt{4/15} = 0.52$. Therefore, we correct the averaged M_0 by adding 0.2 to M_{wp} . Finally, we apply the Whitmore et al. (2002) magnitude-dependent correction $M_{wp} = (M_{wp} - 1.03)/0.843$ to get a final value for M_{wp} (the green traces in Fig. 4).

Figure 5 compares our final M_{wp} values resulting from this procedure (using vertical component velocity data from STS1 and/or KS54000 broadband seismometers for 291 earthquakes of M_w 4.5 to 9.2, occurring between 2002 and 2014) with (GCMT) moment magnitude M_w estimates. For some earthquakes, such as the complex M_w 8.4 (GCMT) Peru event of June 23, 2001, or the great M_w 9.2 (Ammon et al. 2005; Park et al. 2005; Stein and Okal 2007a, b) Sumatra earthquake of December 26, 2004, M_{wp} (7.4 and 8.1, respectively) will underestimate M_w , when the first moment

release is not the largest and/or significant moment release follows the initial moment release by a sufficiently long time. The 2001 Peru M_w 8.4 earthquake began with an initial rupture of moment magnitude approximately 7.4, followed almost 60 s later by a much larger earthquake, of moment magnitude just under GCMT M_w 8.4. In contrast, the PTWC's final estimate of M_{wp} 8.4 for the M_w 8.6 (Centroid Moment Tensor 2008) Nias event of March 28, 2005, was acceptable, as it was for the M_w 8.1 Chile earthquake of April 1, 2014 (Bryant 2001), and for other earthquakes in the M_w 8.0 to 8.4 range. M_{wp} thus enables Regional tsunami warnings (for coastal populations within 1000 km of the epicenter) within 3–5 min of earthquake origin time, as the threshold for a regional warning is M_w 7.6.

At the PTWC, beginning in early 2011, we have automated the M_{wp} method in the following way:

1. Compute a signal-to-noise ratio by examining 60 s of trace prior to, and after, the P-wave



Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings, Fig. 5 A scatter plot of event average PTWC M_{wp} values against GCMT M_w for 291 earthquakes in the M_w 4.5–9.2 range occurring between 1992 and April of 2014. These M_{wp} event averages were derived only from very broadband vertical

seismograms (STS1 and KS5400 seismometers only). The red line represents $M_{wp} = M_w$, and the vertical magenta line defines the NOAA/NWS Tsunami Warning Center's (TWC's) threshold for issuing a local tsunami warning for coastal populations within 300 km of the event's epicenter

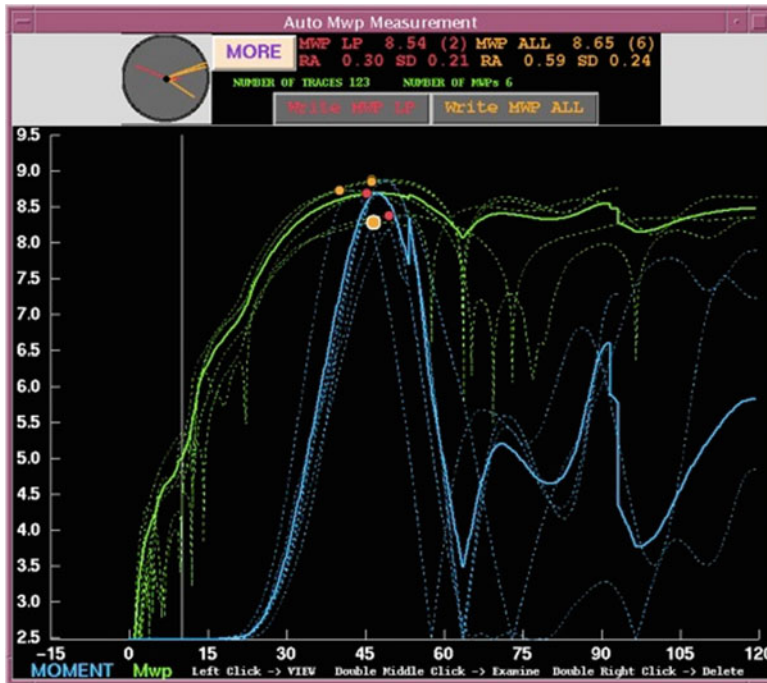
arrival time. Both sections are band-pass filtered between .3 and 5 Hz and the rms mean computed for each. The ratio of these two means yields the signal-to-noise ratio. If the ratio is above 5.0, the auto-Mwp proceeds.

2. Compute the average of the unfiltered noise over a 300 s interval, ending 60 s prior to the P-wave arrival time. This provides a baseline correction for the integration.
3. Doubly integrate the first 120 s of the unfiltered seismogram (minus the noise average determined in step 2), beginning at the P-wave arrival time up to the S-wave arrival or 120 s, whichever comes first, to compute $M_0(t)$.
4. Examine the curve for peaks. This is called the monotonicity check. If there are no clear peaks, then this station may be rejected for auto-Mwp; however it remains available for manual analysis.
5. Recursively smooth the moment(t) curve until three main peaks are left. The algorithm then

uses the location of these peaks as a guide to search for the peaks on the original, unsmoothed moment rate curve. This allows us to apply the Tsuboi method (Tsuboi et al. 1999) to the first two peaks of a smoother waveform, so that this can be done automatically, in real time.

PTWC scientists have also designed a GUI that allows the duty scientists to quickly evaluate the quality of each of these automatically generated M_{wp} measurements.

The GUI (Fig. 6) shows the auto-Mwp results available for review at 6 min after origin time, for the closest six stations to the first of two great earthquakes offshore Sumatra that occurred on April 11, 2012. Note that the resulting M_{wp} value of 8.65 agrees well with the GCMT M_W of 8.6. At the top of the GUI are two summaries, showing the mean M_{wp} based on long-period instruments (red), based on all instruments (orange), and a plot



Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings, Fig. 6 The PTWC's AutoMap graphical user interface (GUI) showing the $M_0(t)$ curves (dashed-blue), the $M_{wp}(t)$ curves (dashed-green), and the stacked $M_0(t)$ and $M_{wp}(t)$ curves, thick blue and green curves, respectively, for a single earthquake. The

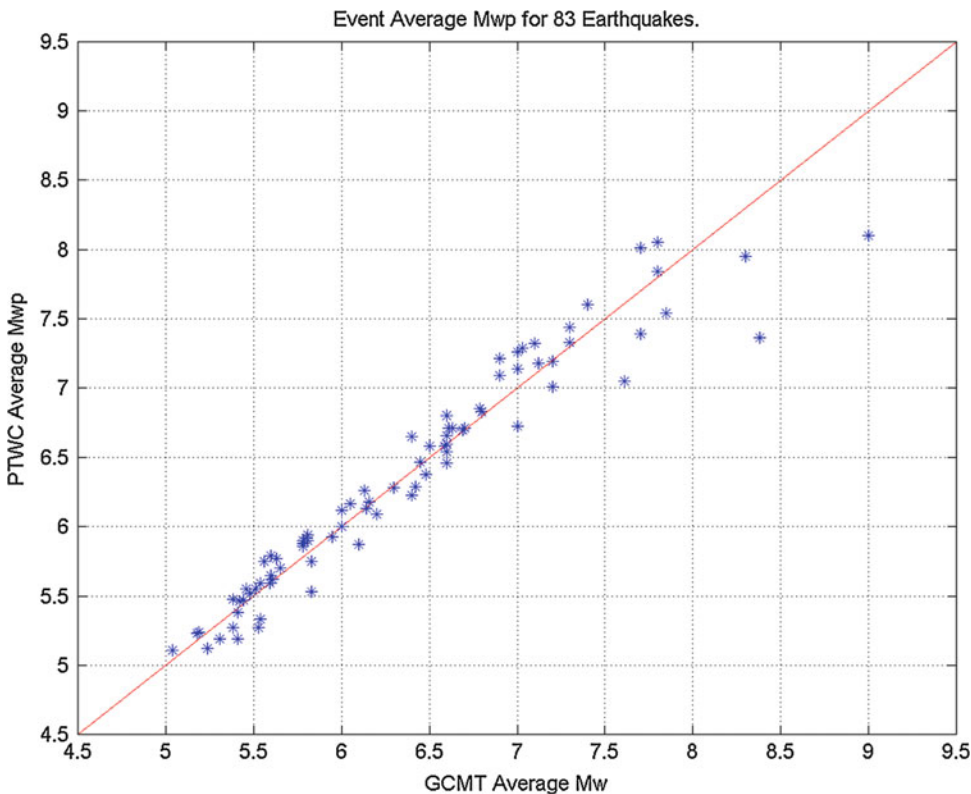
dots show where the auto- M_{wp} algorithm has computed the M_{wp} for the various stations. The stacked $M_0(t)$ curve has one main peak that is populated by six measurements. The solid curves result from stacking the results of all six instruments. The stacked $M_{wp}(t)$ curve (green) is obtained from the stacked $M_0(t)$ curve (blue)

showing the azimuthal distribution of stations, the number of traces processed, and the total number of M_{wp} measurements obtained. There are also three buttons, two of which allow the duty scientist to select the M_{wp} values based on just the long-period instruments or those based on all instruments, and another button “MORE” which will display additional M_{wp} measurements. Auto-Mwp will stop updating M_{wp} in its real-time mode after computing 40 measurements. This is generally many more observations than needed for an accurate estimate of M_W .

The plot (of Fig. 6) shows all of the $M_0(t)$ curves obtained (dashed-blue), the $M_{wp}(t)$ curves (dashed-green), and the stacked $M_0(t)$ and $M_{wp}(t)$ curves, thick blue and green curves, respectively. The dots show where the auto- M_{wp} algorithm has computed the M_{wp} for the various stations. As can be seen in Fig. 6, the

stacked $M_0(t)$ curve has one main peak that is populated by six measurements. The solid curves result from stacking the results of all six instruments. The stacked $M_{wp}(t)$ curve (green) is obtained from the stacked $M_0(t)$ curve (blue).

The PTWC also calculates M_{wp} from locally recorded P-waves to estimate M_W for large ($>\sim$ Mw 5) local earthquakes, occurring in the Hawaiian Islands (Hirshorn 2004, 2007), Puerto Rico, the Virgin Islands, and California. Because M_{wp} is based on the far-field approximation to the (vertical) P-wave displacement due to a double couple point source (Tsuboi et al. 1995), we should only very carefully apply M_{wp} to data obtained in the near field, at distances of less than a few wavelengths of the fault. At about 5 degrees epicentral distance, the point source approximation appears to be valid up to about GCMT M_w 7.5–8 (see Fig. 7). Figure 7 is a plot of M_{wp} values calculated from locally recorded



Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings, Fig. 7 A scatter plot of 83 Mwp event averages obtained from observations

within 5 degrees epicentral distance from their respective epicenters, plotted against the GCMT event Mw for the same earthquakes

P-waves, up to 5 degrees epicentral distance, against the GCMT M_W values for the same events. For example, PTWC scientists calculated a value of M_W 6.5 for the GCMT M_W 6.7 Kiholo Bay, Hawaii, earthquake within 3 min of the initiation of rupture at its hypocenter (Hirshorn 2007). Based on data provided later to the PTWC for the 2011 GCMT 9.1 Off Tohoku earthquake, accurate M_{wp} determinations of 9.3, 9.3, and 9.1, were obtained at 10.9, 12.4, and 14.9 degrees epicentral distance, respectively (Hirshorn et al. 2013). M_{wp} estimates from these 18 STS-1 broad band velocity records at 2.2–14.9 degrees epicentral distance, all exceeded the PTWC's threshold for a Tsunami Threat Message for coastal populations within 1000 km of the earthquake.

The Mantle Magnitude (M_m) Method

Emile Okal and J. Talandier developed the M_m method in 1989 (Okal and Talandier 1989). The mantle magnitude is related to the moment magnitude via the simple expression $M_W = M_m / 1.5 + 2.6$. This work was inspired by the need to develop a magnitude method for tsunami warning centers that would not suffer the saturation problem that M_S suffers (Okal 1992a). Not only may M_S saturate as M_W becomes large (>8), but slow earthquakes can also cause M_S to be seriously deficient and $bMag$ even more so. Severely underestimating the magnitude of an earthquake can lead to a failure to warn. PTWC's implementation of the M_m method is based on analyzing Rayleigh waves obtained on vertical component seismograms.

M_m , being based on slow traveling long-period surface waves, is available too late to be used in the decision process for issuing an initial bulletin. Notwithstanding, it does provide a useful check on the magnitude obtained from the M_{wp} method, and if there is a discrepancy between M_{wp} and $M_W(M_m)$ on the order of 2–3 tenths or more in the 7⁺ magnitude range, then the duty scientist may instead use the results of the M_m method in subsequent bulletins. The M_m method overcomes the limitation of saturation because it is a variable period magnitude. Multiple values of M_m are routinely computed for a number of fixed periods ranging from 50 to 270 s for each station. Because

M_m may saturate at smaller periods for great earthquakes while remaining unsaturated at longer periods, Okal and Talandier's (1989) procedure was to choose the largest M_m , thus mitigating the effects of saturation.

M_m is more complicated than the other magnitude methods described here as it uses frequency-domain deconvolution. This can cause problems due to deconvolution noise at low magnitudes, where the amplification of noise by the deconvolution process at long periods may result in spurious magnitudes. Thus M_m works best with very long-period broadband seismometers. While STS-2 seismometers tend to do well, the shorter-period broadband seismometers tend to behave poorly at the longest periods (Weinstein and Okal 2005). Using the maximum M_m obtained for each station proved to be suboptimal due to the heterogeneous distribution of instruments coupled with the total automation of the procedure at the PTWC. Weinstein and Okal (2005) devised a sampling method that alleviates most of these difficulties in PTWC's implementation of M_m .

The December 2004 Sumatra earthquake showed that for earthquakes with an unusually long source duration (in this case ~600 s) (Ishii et al. 2005; Kanamori 2006), even M_m calculated using 270 s waves will saturate. Hence PTWC's M_m implementation will now automatically extend the period range to 410 s when the magnitude exceeds 8.0. At 410 s $M_W(M_m)$ is 8.9 (Weinstein and Okal 2005) for the GCMT M_W 9.2 December 2004 Sumatra earthquake, still deficient, but a marked improvement over the moment magnitude 8.5 obtained by PTWC and 8.2 obtained by the USGS (NEIC Fast Moment Tensor) on Dec. 26, 2004. M_m normally uses a 660 s window of the surface wave train, but when M_m exceeds 8.0, it expands this window to 910 s. Given the mix of instruments and their distribution used by the PTWC and the effects of broadband deconvolution noise, Weinstein and Okal (2005) found that the M_m method was usually not useful for earthquakes smaller than M_W 6.0. Since publication of that study, PTWC has made some refinements to the M_m calculation. PTWC computes the "noise floor," i.e., the M_m spectra using data obtained before the earthquake origin

time, for each seismic station. M_m values obtained for the earthquake are compared to those obtained from the pre-event noise. If the seismic moment obtained at any period is less than a factor of 10 above the seismic moment obtained from the pre-event noise, the value for M_m at that period is rejected. For smaller earthquakes with $M_w < 6.0$, this typically means that values of M_m at longer periods get rejected. With this technique, PTWC routinely obtains credible magnitude estimates based on mantle waves for earthquakes with magnitudes as small as 5.5 and sometimes even smaller depending on the noise floor. The same technique also eliminates measurements that may be contaminated by mantle waves generated by very recent earthquakes.

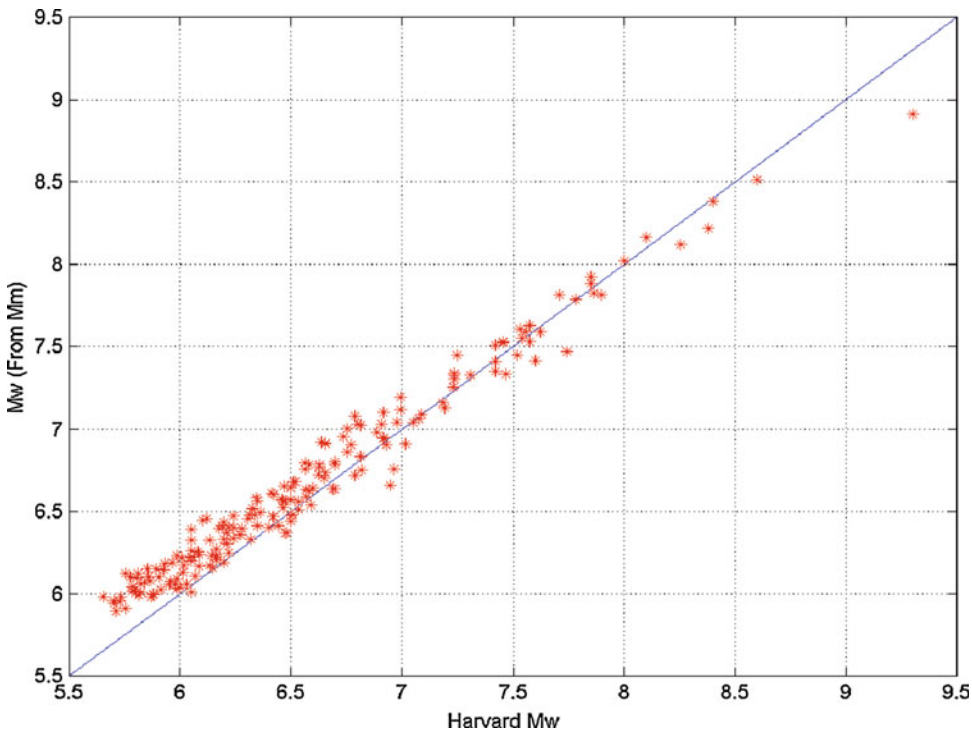
Figure 8 compares $M_W(M_m)$ values obtained for more than 200 earthquakes with the respective GCMT M_W values for the same events. PTWC's implementation of M_m tends to overestimate M_W by 0~0.15 magnitude units for $M_m < 7.0$.

The W-Phase Method

H. Kanamori first identified the W-phase on displacement seismograms generated by the 1992 Nicaragua tsunami earthquake (Kanamori 1993). This earthquake was deficient in high-frequency radiation but rich in long-period energy (see Fig. 9).

When the periods of phases such as P, PP, S, SS, SP, etc. approach or exceed the travel time differences between them, the phases will interfere with each other. At these long periods, the result of this interference is itself a distinct phase now called the W-phase (Kanamori 1993).

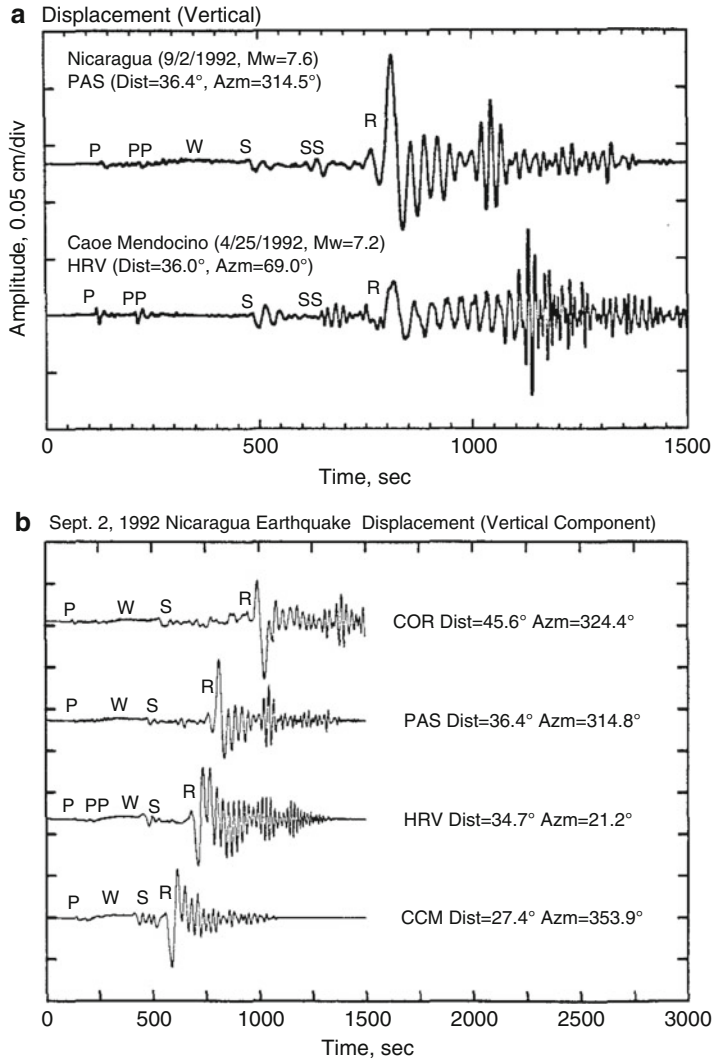
The name W-phase was chosen because this phase is analogous to the "whispering gallery" mode of wave propagation found in acoustics or optics (Kanamori 1993). An example of the optical whispering gallery mode is light trapped inside a raindrop. The raindrop acts a like an optical fiber loop with the light strongly concentrated near the surface of the drop. P. Cummins showed that the W-phase could also be modeled



Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings, Fig. 8 Scatter plot of $M_w(M_m)$ vs. GCMT M_w for over 200 earthquakes

Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings,

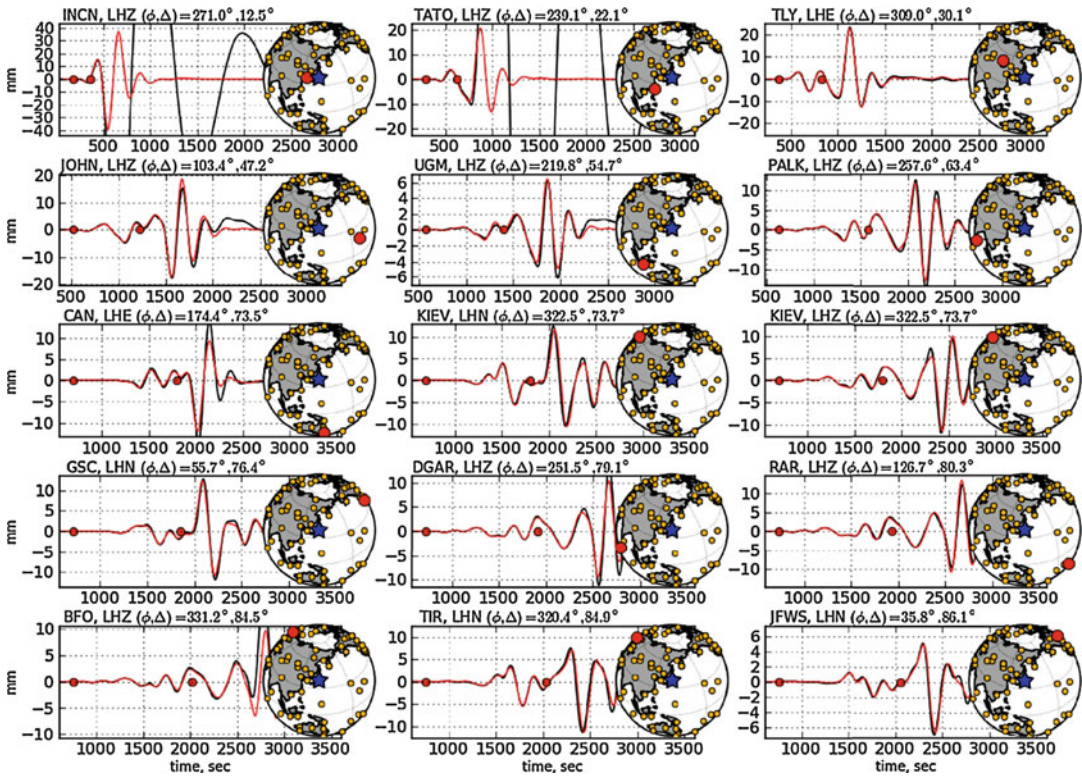
Fig. 9 W-phase observed for the September 2, 1992, Nicaragua earthquake. Figure 10a compares the Nicaragua earthquake with a Cape Mendocino earthquake. Figure 10b shows the seismograms of the Nicaragua earthquake recorded at stations COR, PAS, HRV, and CCM. All the seismograms show displacement computed from the original broadband seismograms. (Figure 1 from Kanamori 1993)



as the complete elastic wave field (near field + far field) (Cummins 1997).

In the context of normal mode theory, the W-phase can be interpreted as the superposition of the fundamental mode and the first, second, and third overtones of spheroidal modes at long periods (Kanamori and Rivera 2008). The group velocity of the W-phase varies from 4.5 km/s to 9.0 km/s over a period range from 100 to 1000 s. Like mantle waves, most of the energy in the W-phase propagates within the mantle, so that the propagation of the W-phase is little affected by the complications of structural heterogeneities such as continents and oceans.

Because of the W-phase's very long-period character, magnitude estimates based on it should not saturate, even for truly great earthquakes. H. Kanamori and L. Rivera demonstrated this for the great M_w 9.2 Sumatra earthquake of 2004, with a source duration of ~600 s (Kanamori and Rivera 2008). The true magnitude of the Sumatra earthquake of 2004 was not known until 2005, when enough seismic data were recorded (weeks) for stacking, to allow the amplitudes of the Earth's free oscillations to be accurately measured (Park et al. 2005). H. Kanamori and L. Rivera's W-phase method (Kanamori and Rivera 2008) now provides magnitudes and centroid moment



Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings, Fig. 10 Examples of observed waveforms (black lines) and the corresponding synthetics (red lines) computed from the W-phase solution for the GCMT Mw 9.1 Tohoku earthquake. The station azimuth (ϕ) and epicentral distance (Δ) are indicated as

well as W-phase time window, bounded by red dots. W-phase and later arrivals are often very well predicted by the solution. For some channels like INCN-LHZ or TATO-LHZ, the surface waves are affected by instrument problems, though the earlier-arriving W-phase signal is not affected. (Figure 3 from Duputel et al. 2011)

tensors for great earthquakes to PTWC duty scientists, within about 25 min after the initiation of rupture at the hypocenter. For the 2011 M_w 9.0 Tohoku, Japan, earthquake, PTWC's W-phase implementation yielded an M_w of 9.0 when provided with the correct depth (Duputel et al. 2011) (Figs. 10 and 11).

Luis Rivera, Zacharie Duputel, and PTWC scientists implemented the W-phase method at the PTWC in late 2010. W-phase processing begins when the PTWC issues an observatory message. If the earthquake's magnitude is ≥ 5.8 , then the software will compute a W-phase CMT. W-phase results are sent to the duty scientists if the magnitude in the observatory message is 6.8 or larger. The W-phase CMT computations for earthquakes in the M_w 5.8 to 6.7 range enable research and debugging, but are not used for tsunami warning purposes.

The GCMT M_w 9.1 Tohoku earthquake of March 10, 2011, provided an excellent test of the PTWC's W-phase processing. Figure 11 is a table showing the evolution of the W-phase CMT with time (Duputel et al. 2011). The PTWC obtained an initial W-phase CMT 22 min after the earthquake's origin time. Because our initial depth determination was too deep, the magnitude was underestimated. When manually re-triggered with a better hypocentral depth, 44 min after origin time, PTWC obtained M_w 9.0 and a CMT consistent with the USGS and GCMT M_w values (Hayes et al. 2011).

Tsunami Earthquakes

Tsunami earthquakes are so-called because they generate tsunamis that are more destructive than expected given their magnitude. Classic examples include the Nicaragua 1992 (Fig. 9) and Java 2006

Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings,
Fig. 11 Real-time W-phase results from the USGS and the PTWC from Duputel et al. (2011)

4. Real time results Tohoku-oki 11-03-2011 (Mw=9.0)						
Delay	Origin	M _w	Dip, deg	Depth, km	# chan.	Mech.
20min	USGS	9	16.8	24.4	6	
22min	PTWC	8.8	10.3	83.5	29	
30min	PTWC	8.8	22.8	83.5	74	
40min	PTWC	9	11.1	24.4	105	
45min	IPGS	9	10.8	24.4	31	
48min	USGS	8.9	14.8	24.4	74	
1h02min	USGS	8.9	16.9	24.4	89	
1h30min	IPGS	9	14.4	24.4	146	

earthquakes. Tsunami earthquakes present a challenge to tsunami warning systems because they generate much less ground shaking than expected given their size. Hence populations near the earthquake, which otherwise might be alerted by strong ground shaking, may not self-evacuate if they are not alarmed by the ground shaking. Tsunami or “slow” earthquakes also present a challenge to warning centers because their unique rupture characteristics affect magnitude determinations. These two characteristics, i.e., difficulty in determining correct magnitude and generating larger tsunamis than expected (even with the correct magnitude), contribute to the possibility of under warning coastal populations of a destructive tsunami. Fortunately, there are methods to alert duty scientists to the occurrence of tsunami earthquakes.

One way in which the occurrence of a tsunami earthquake may be detected is if the bMag and/or MS are significantly smaller than Mwp obtained from the longer-period P-waves or longer-period mantle waves. Figure 12 shows this clearly for short-period $\hat{m}_b$ (Gutenberg and Richter 1956b) Vrs. M_w . Note the population of four tsunami earthquakes that fall well off the trend. As can be determined from Fig. 12, the body-wave

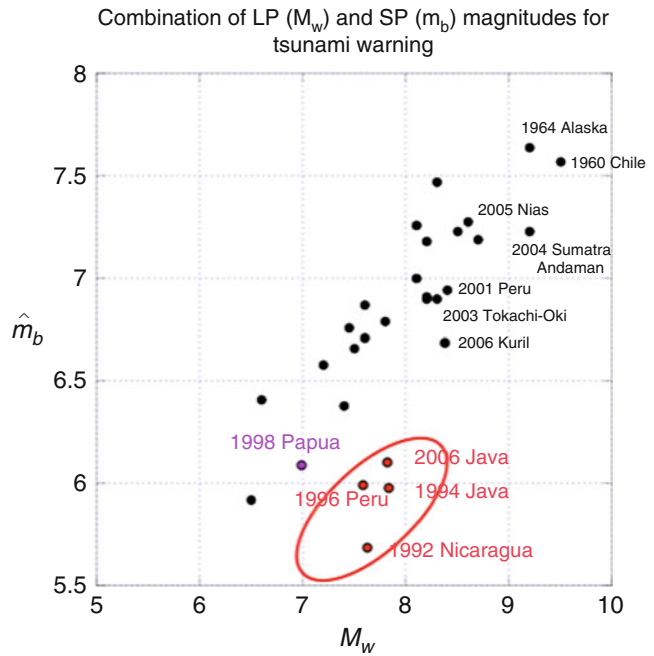
magnitude for July 2006 Java earthquake was deficient by nearly 1.5 magnitude units.

As mentioned earlier, a fundamental characteristic of a tsunami earthquake is the slowness of the rupture speed. Newman and Okal (1998) showed that the log ratio of the radiated energy E_R (Boatwright and Choy 1986; Boatwright et al. 2002), to the seismic moment M_0 , $\text{Log}_{10}(E_R/M_0)$ (also denoted by theta or “ θ ”) is anomalously small for tsunami earthquakes. A number of factors can affect this ratio such as rupture velocity, stress drop/apparent stress, fault plane geometry, maximum strain at rupturing, and directivity (bilateral vs. unilateral rupture). However, for shallow thrust, low stress drop subduction zone earthquakes, unusually slow rupture velocity may have the largest influence on the value of θ .

Newman and Okal (1998) showed that for tsunami quakes, the value of θ is usually -6.0 or less. For an earthquake with a unilateral rupture and nominal speed (~ 3 km/s), theory suggests that θ is -4.9 (Geller and Kanamori 1977; Scholz 1982; Vassiliou and Kanamori 1982). Weinstein and Okal (2005) extended the original dataset of Newman and Okal (1998) by including an additional 118 earthquakes. The mean value of all θ values is approximately -5.1 . However, when

Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings,

Fig. 12 Comparison between short-period $\hat{m}_b$ (Gutenberg and Richter 1956b) and M_w for earthquakes with $M_w > 6$. Note the cluster of red dots representing tsunami earthquakes. This illustrates the diagnostic potential of short-period/long-period magnitude ratios to identify unusually slow earthquakes with high tsunami potential (This is Hiroo Kanamori's figure from his talk at the PTWC in April of 2007)



averaged by event, the distribution of θ 's peaks precisely at -4.9 , in accordance with theoretical expectations. Given the standard deviation of 0.39 for all θ 's (for the 118 earthquakes), Weinstein and Okal (2005) found that values of θ around -6.0 or below are more than 2σ off of the mean and hence clearly anomalous.

The PTWC uses broadband vertical component seismograms, obtained in the distance interval $5^\circ \leq \Delta \leq 90^\circ$, to compute θ . Originally, the closest distance was 25° . However, this can be made as close as 5° using new corrections (Ebeling et al. 2012). A window of 75 s is used starting approximately 5 s before the P-wave arrival to insure that the first arrivals are not missed by the integration. This window is deconvolved with the instrument response, and the radiated energy contained between 1 and 2 Hz is computed.

In general, it is thought that anomalously slow rupture speed is due to either low-rigidity sediments in the fault and/or faulting through an accretionary prism (Bilek and Lay 1999; Bilek et al. 2004; Fukao 1979; Kanamori 1972; Kanamori and Kikuchi 1993; Satake 1994). In either case, the small shear rigidity associated with weak materials retards the rupture speed.

As to why "slow" earthquakes produce more destructive than expected tsunamis, one can look at the well-known relation for moment magnitude:

$$M_0 = \mu A d$$

where μ is the shear modulus, A is the fault plane area, and d is the average slip over the fault plane. Thus, for two earthquakes with the same seismic moment (M_0), and fault area (A), lowering μ (i.e., decreasing the rupture propagation speed) requires increasing the slip (d) to keep the M_0 (and M_w) the same. Hence a slow rupturing, Tsunami earthquake produces larger-than-expected slip and therefore tsunami, based on seismic moment.

One problem with θ is that it can be misleading and occasionally yield false indications of rupture slowness. This was made apparent by the Peru earthquake of June 23, 2001. This earthquake began with an initial event that had a moment magnitude of approximately 7.4, followed almost 60 s later by a much larger event, which had a moment magnitude of almost 8.4. (Bilek and Ruff 2002; Giovanni et al. 2002; Kikuchi and Yamanaka 2001). Due to the 60 s delay, the θ

Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings, Fig. 13 The variation of Θ with offset for (a) a “normal” earthquake, (b) a “slow earthquake” (Java, 2006), and (c) a complex earthquake (Peru, 2008), respectively. In these plots Θ is demeaned (the mean is found on the bottom right of the plot in black), and the number next to the dots indicates how many stations were used in computing that value. These plots are taken from PTWC’s operational software



computation used mainly P-wave coda from the first shock and little if any energy from the main shock. As a result, the PTWC initially obtained a θ of -6.1 , using a moment based on M_m , making this earthquake appear very slow indeed. However, this result is spurious and was due to the complexity of the earthquake itself and not to actual slowness of the rupture.

Weinstein and Okal (2005) found that by sliding the window over which θ is computed forward in time, θ would increase as the θ window overlapped with the occurrence of the main event of the Peru 2001 earthquake. Indeed, for a window offset of 70 s, θ increases to -5.6 , which is a strong trend to slowness, but not a slow or tsunami quake. This was further borne out by the size of the tsunami, which while detected on sea-level instruments around the Pacific (more than 2 m peak to trough in Chile) was not destructive outside of Peru.

Weinstein and Okal (2005) explored the windowing technique and found that it was a more comprehensive method than the single

determination of θ (zero offset). Computing theta in a succession of windows separated in time by 10 s (each window spanning 70 s) up to 100 s post P-wave arrival yields a better method of detecting slowness; see Fig. 13. What Weinstein and Okal (2005) found is that for true tsunami earthquakes, the variation of θ with offset time was small, generally no more than 0.1 log units over the entire 100 s. It is this flat trend that is probably the best discriminant for tsunami earthquakes.

In effect, the curve resulting from the window-offset technique gives an indication of the source duration of the earthquake. Gigantic earthquakes have long source durations, and slow earthquakes have anomalously long source durations for their seismic moment; see Fig. 13. Therefore θ can be viewed as a measurement of how anomalous the source duration is in terms of whether the earthquake is anomalously slow or simply anomalously large. It turns out that in the case of the Sumatra earthquake of December 2004, θ has little variation, even when the integration window

is increased to 200 s and the offset carried out to 300 s. The magnitude of θ based on PTWC's M_w (Mm) of 8.5 was ~ -5.6 , a trend to slowness, but not slow. Using the M_w based on normal mode studies, θ is ~ -6.1 (with a 200 s integration window!), and discussion continues to the current day as to whether or not the Sumatra earthquake of 2004 was slow and simply had aspects of a tsunami earthquake or none at all (Ammon et al. 2005; Global Centroid Moment Tensor (GCMT); Krüger and Ohrnberger 2005; Lay et al. 2005; Seno and Hirata 2007).

Real-Time Tsunami Forecasting

Historically, tsunami warning centers (TWCs) issue tsunami warnings based on the initial earthquake magnitude (based on historical events) and tsunami travel times. Pre-computed databases of tsunami scenarios can also be used to provide a quick estimate of a possible tsunami threat (Gica et al. 2008; Kamigaichi 2009; Kowalik and Whitmore 1991; Tatehata 1997). The principle limitation of the database approach is that it cannot exhaust all possible earthquake source parameters and locations. Due to advanced computer technology and the development of rapid real-time source inversion methods such as the W-phase method (Duputel et al. 2011, 2012; Hayes et al. 2009; Kanamori and Rivera 2008), real-time tsunami forecasting, using real-time earthquake parameters and other contemporaneous observations (e.g., GPS and water level observations), has become a reality. W-phase centroid moment tensors (CMT) obtained during events have been used in a real-time forecast model at PTWC (Duputel et al. 2011; Fryer et al. 2010; Wang et al. 2012) since 2011. Offshore GPS buoy data can be used to constrain the tsunami sources in real time (Yasuda and Mase 2013). Land-based GPS data can also be used to constrain the tsunami source in a real-time model (Hoechner et al. 2013).

PTWC started experimenting with a real-time tsunami forecasting model RIFT in 2009. Starting in 2014, RIFT became the basis of the forecast products issued with PTWC tsunami threat messages (Wang et al. 2009). The typical work cycle is the following:

- (1) As soon as the earthquake has been located and the magnitude determined, the model is run for regions near the epicenter (within several hours of tsunami arrival time), using the real-time earthquake location and magnitude with default focal mechanisms. The default focal mechanism is based on the type of plate boundary the earthquake is closest to. To be conservative, a pure thrust mechanism along subduction zones can be assumed, and the strike used is the strike of the trench axis. Alternatively, a normal faulting or strike-slip faulting focal mechanism can also be used if the epicenter is near a divergent plate boundary or a strike-slip plate boundary, respectively. The model results for the near field can be obtained in a few seconds.
- (2) Once the W-phase centroid moment tensor becomes available (typically within 20–30 min of the earthquake origin time), the model is run again to update the forecast. A larger model domain (e.g., the entire Pacific basin) can be used for large earthquakes.
- (3) The model forecast is compared with water level observations to confirm or refine the forecast.

The RIFT model is based on the finite difference discretization of the linear shallow water equations:

$$u_t = -g\nabla\eta,$$

$$\eta_t + \nabla \cdot (uh) = 0,$$

where $u = (u, v)$ is the vertically averaged horizontal velocity, η the sea surface elevation, h the ocean depth at rest, and ∇ the horizontal gradient operator. The model is discretized in spherical coordinates with a leapfrog time differencing scheme on the Arakawa C-grid (Arakawa and Lamb 1977). The numerical scheme is similar to that of Imamura except that the staggered leapfrog time differencing scheme is used (Imamura 1996).

The initial condition is based on the static seafloor deformation formula of Okada (1985) for a rectangular rupture zone with a uniform slip. The assumed rupture size is a continuous

function of earthquake magnitude (Henry and Das 2001; Wells and Coppersmith 1994). The coastal wave amplitude is based on the Green's law (Lamb 1932),

$$a_c = a_o(h_o/h_c)^{1/4}$$

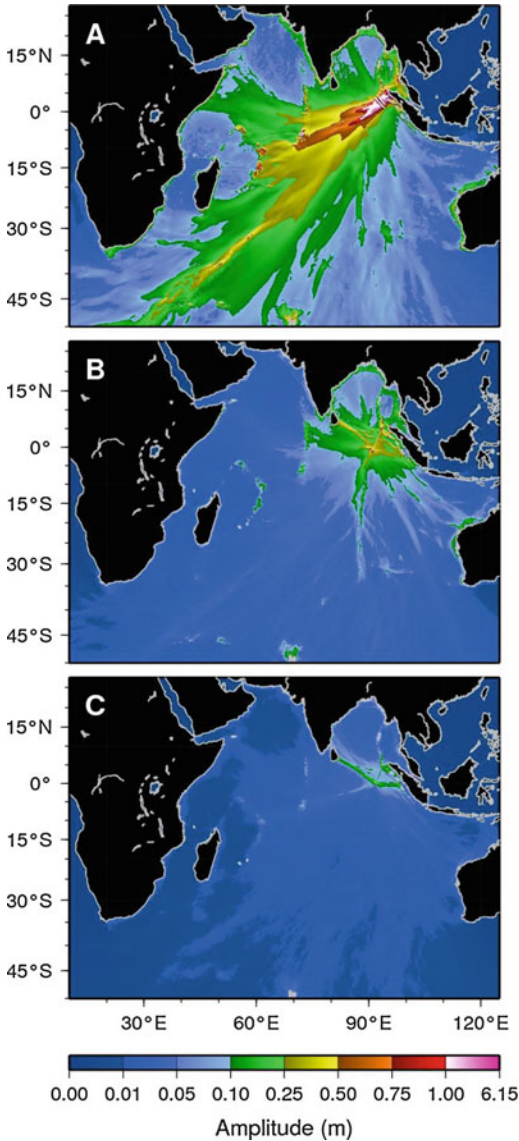
where a_c and a_o are wave amplitudes at a coastal point and an corresponding offshore point. h_c and h_o are water depths at the coastal point and the offshore point, respectively. The offshore point is chosen such that the water depth is greater than a threshold value (typically 1000 m at 4-arc-min resolution or smaller depth for finer resolution). The wave amplitude is defined as half of the wave height, which is the difference between the trough and crest of the wave. We should point out that the Green's law coastal forecast is most applicable for coastlines that are directly exposed to the open ocean. The Green's law is not applicable for locations that are not exposed to the open ocean directly (such as harbors well hidden from the open ocean, fjords, and rivers). Nevertheless, the Green's law forecast should provide a quick order of magnitude estimate of general tsunami threat for tsunami warning purposes.

The bathymetry used by RIFT is based on the 30-arc-sec resolution GEBCO data (Becker et al. 2009). The operator of the model can choose 1 of 40 predefined ocean basins or use a basin determined on the fly by a tsunami travel time software (Wessel 2009; Wessel and Smith 1991). This is the default basin setting. A forecast for this default model domain can be obtained in ~ 10 s at 4-arc-min resolution on a 12-CPU Linux server. A 30-h forecast for the entire Pacific basin at 4-arc-min resolution can be obtained in 7 min. For smaller earthquakes (magnitudes < 7.5) or earthquakes that only have a regional impact (e.g., a local earthquake in Hawaii), a smaller domain and higher resolution can be used.

Timeliness is essential for tsunami warning. The linearization of the shallow water equations and the use of the Green's law in the RIFT model allow for rapid forecast of tsunami threats. The RIFT model was run for most of the events since 2009 for which measurable tsunamis were recorded. The results are encouraging. As an

example, we discuss the April 11, 2012, magnitude 8.6 earthquake off the northern Sumatra coast and the ensuing tsunami (Wang et al. 2012). Based on an initial earthquake magnitude of 8.7, the PTWC issued an Indian Ocean basin-wide tsunami watch. The governments of India, Indonesia, Sri Lanka, Thailand, and the Maldives also issued Tsunami warnings.

Figure 14 shows three RIFT forecast results obtained during the event. The top panel (Fig. 14a) shows the maximum wave amplitude for a default shallow thrust mechanism (based on PTWC's initial earthquake Mwp magnitude of 8.66). The forecast was obtained 14 min after the earthquake origin. It forecasts a tsunami posing a significant threat to some Indian Ocean countries, with a maximum coastal wave amplitude using Green's law of 9 m. The middle panel (Fig. 14b) shows the maximum wave amplitude from the RIFT model using the PTWC's W-phase CMT solution ($M_w = 8.78$) obtained 39 min after the earthquake origin time. Although the earthquake magnitude obtained from this CMT was greater than the initial estimate based on the M_{wp} method, the tsunami wave amplitudes were reduced substantially. The maximum coastal wave amplitude was only 2.4 m near the epicenter, and only a handful of coastal locations experienced wave amplitudes over 1 m. This reduced tsunami amplitude is due to the earthquake's strike-slip focal mechanism. The bottom panel (Fig. 14c) shows the maximum wave amplitude for the RIFT model solution using the USGS W-phase CMT ($M_w = 8.57$), obtained 1 h 40 min after the earthquake's origin time. The maximum forecasted coastal wave amplitude barely reached 1 m. This forecast combined with the sea-level observations allowed the PTWC to cancel the basin-wide tsunami watch sooner than would have been possible without the forecast. This event demonstrates the usefulness of obtaining an accurate focal mechanism for tsunami forecasting and the associated problems with magnitude-based tsunami warnings. Although we cannot yet avoid magnitude-based warnings for near field coastal populations, knowing the focal mechanism of the earthquake allows more accurate tsunami forecasting for the far field.



Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings, Fig. 14 Energy (maximum wave amplitude) maps for the April 11, 2012, Northern Sumatra tsunami. (a) Forecast based on PTWC's initial $M_w = 8.66$, (b) forecast based on PTWC W-phase CMT ($M_w = 8.78$), and (c) forecast based on USGS W-phase CMT ($M_w = 8.57$)

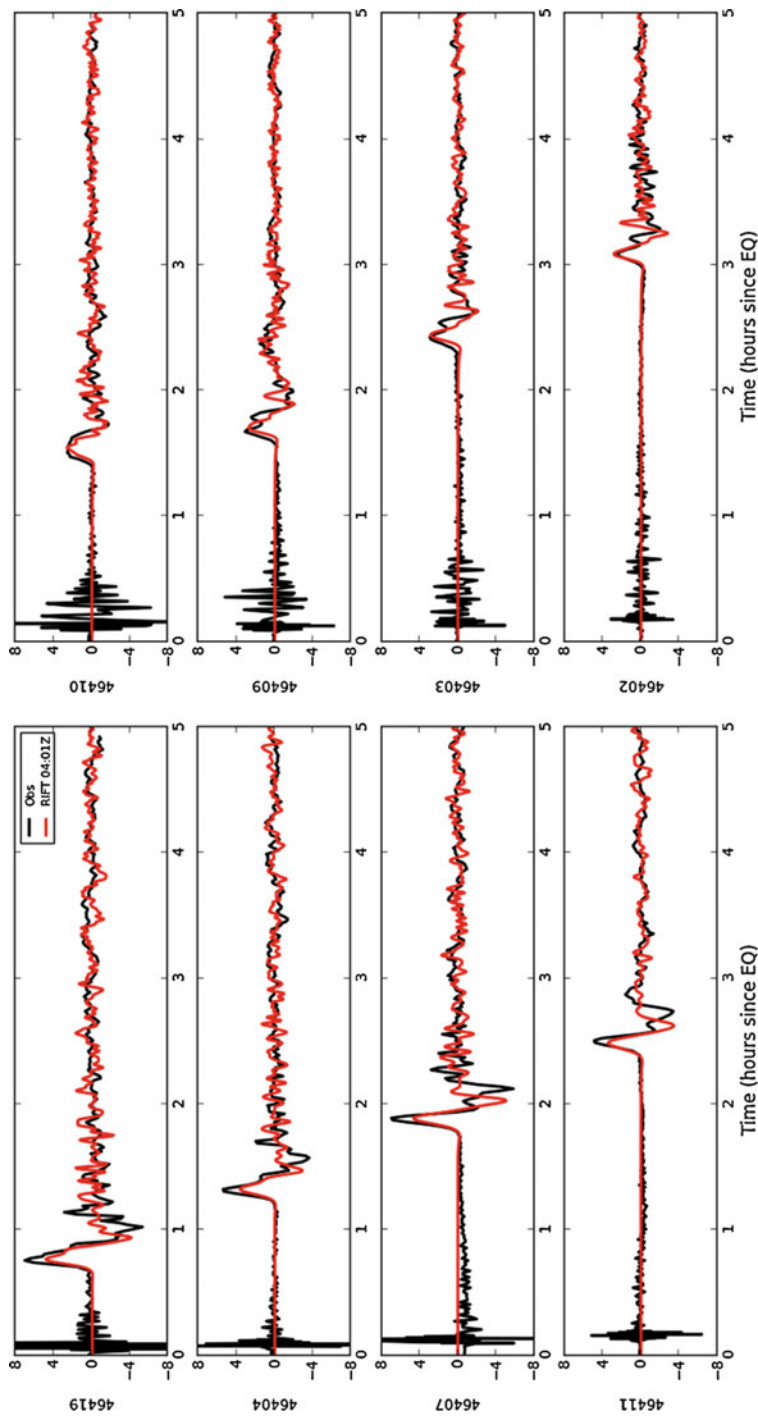
The October 28, 2012, 3:04 UTC Haida Gwaii, British Columbia, Canada, magnitude 7.7 earthquake provided another example of the usefulness of the CMT derived from the W-phase inversion for real-time tsunami forecasting. Historically, this region has been characterized by strike-slip

earthquakes. This event, however, had a thrust mechanism, as shown by the W-phase CMT solution. Figure 15 shows a comparison of observations at DARTs (deep ocean bottom pressure sensors) with RIFT forecasts obtained 57 min after the earthquake origin time, based on the USGS W-phase CMT ($M_w = 7.7$). The forecast was reasonably good across the basin, although in the near field, the forecast underestimated the wave amplitudes. We note that at the time of the forecast, no DART stations had yet recorded a complete waveform of the tsunami.

Figure 16 compares observations obtained for the Haida Gwaii event from 57 tide stations throughout the Pacific with RIFT forecast generated at 0401Z. Although there is considerable scatter between the observed and predicted wave amplitudes (defined as an average of maximum zero to peak and maximum zero to trough amplitudes), the mean ratio between the predicted and observed wave amplitudes is 1.4, or prediction is within a factor of two of the observed. If only tide stations that are more exposed to the open ocean are included (38 out of 57 tide stations), then the mean ratio is 1.07. We note that tsunami warning cannot be based on forecast at tide stations. For example, during the 2009 Samoa tsunami, post-event survey showed run-up reaching 7 m on the island of Tutuila in American Samoa, but the tide station in Pago Pago Harbor only recorded a maximum wave amplitude of ~ 2 m. In summary, the W-phase CMT method combined with a linear shallow water model and the Green's law can provide useful and rapid guidance for tsunami warning, complementing existing database approaches. Real-time inundation forecasting based on propagation forecast results obtained using the CMT obtained by inversion of the W-phase can also be added in the future.

Future Directions

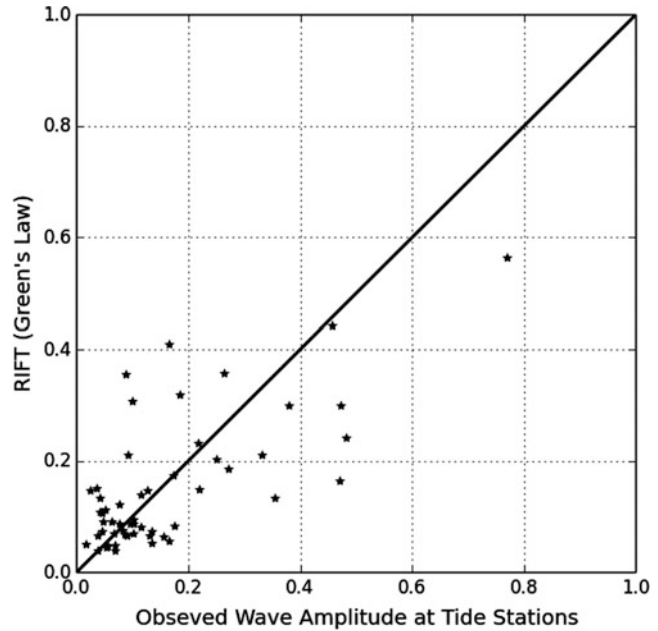
Given the availability of high-quality broadband seismic data, the tsunami warning centers can determine basic earthquake source parameters rapidly. However, the source characterization at the warning centers has rested largely on scalar



Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings, Fig. 15 Comparison of tsunami recorded at DARTs across the Pacific with RIFT forecast based on USGS W-phase CMT ($M_w = 7.7$). The DART buoy locations and data can be found at: <http://www.ndbc.noaa.gov/> (October 28, 2012, Haida Gwaii tsunami)

Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings,

Fig. 16 Comparison of observed wave amplitudes at tide stations with RIFT forecast based the USGS W-phase CMT. Wave amplitude is defined as the average of maximum zero to peak and maximum zero to trough wave amplitudes. (October 28, 2012, Haida Gwaii tsunami)



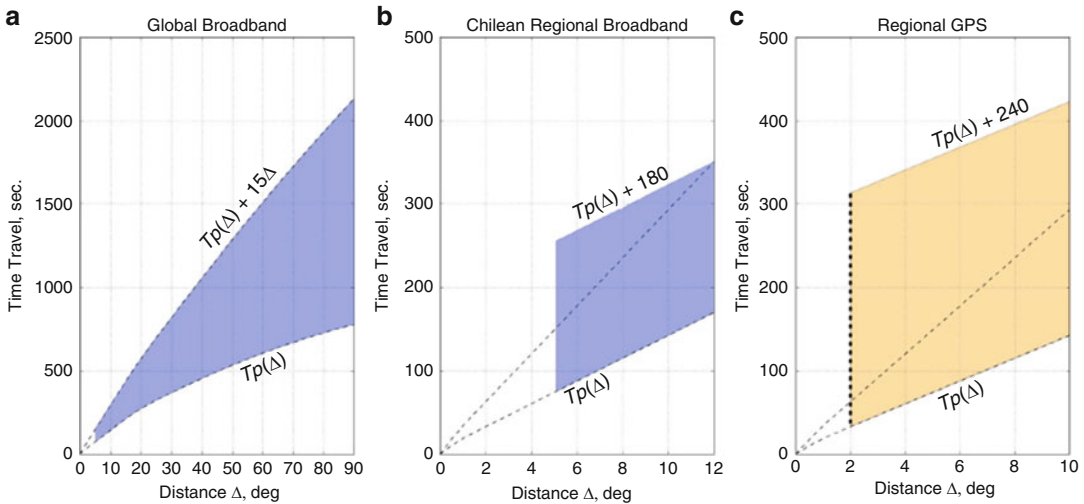
measures of earthquake magnitude. The reasons for this are historical and practical. The warning centers have not always received the quantity and quality of seismic data that they do now, and in the interest of speed, the calculation of scalar measures can be accomplished with the data at hand in a small amount of time. One issue that the PTWC faced during the 2004 Sumatra earthquake was that no near real-time magnitude method existed at the time that would correctly estimate the size of the Sumatra earthquake. Since then, new techniques have been developed to determine the magnitude of great earthquakes. Among these are techniques are the W-phase source inversion method (Kanamori 1993; Kanamori and Kikuchi 1993; Cummins 1997; Kanamori and Rivera 2008).

Luis Rivera, Zacharie Duputel, and PTWC scientists implemented the W-phase method (Kanamori and Rivera 2008) at the PTWC in late 2010, using data from seismic stations in the 5 to 50 degree epicentral range. The first W-phase CMT and unsaturated M_w estimate are therefore available at 25 to 30 min after earthquake origin time.

Using data from the Earthquake Research Institute F-net, for eight large Japanese

earthquakes, Kanamori H. and Rivera L. (Kanamori et al. 2008) showed that, because the W-phase includes the complete wave field, including both near- and far-field displacements (Cummins 1997), it could be used for regional tsunami warnings, reducing its availability for tsunami from approximately 25 to 6 min after earthquake origin time. They achieved this with only minor modifications to the algorithm:

1. Using the teleseismic windowing scheme (see Fig. 17a), the duration of the data becomes too short at short distances. Thus, Kanamori and Rivera (Kanamori et al. 2008) used a constant window with a duration of $15\Delta_0$ where Δ_0 is a fixed distance, 12° . In this case the record duration is 180 s (see Fig. 17b).
2. The records are often clipped very early at short distances; thus, Kanamori and Rivera (Kanamori et al. 2008) removed all of the stations at distances shorter than 5° from their W-phase inversion. Also, they used a maximum epicentral distance of 12° , which provided W-phase-based moment tensor solution 6 min after the earthquake's origin time.
3. For regional applications, one often needs to determine the source mechanism of smaller



Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings, Fig. 17 Time window comparison between (a) teleseismic W-phase inversion from seismic data, (b) regional inversion from seismic data, and (c) regional inversion using GPS data. (a) shows a

window from $T_p(\Delta)$ to $T_p(\Delta) + 15\Delta$, where Δ is the station to centroid distance in degrees from 5° to 90° ; (b) shows a constant window of 180 s with Δ between 5° and 12° distance; and (c) shows a constant window of 240 s for Δ between 2° and 10° . (Fig. 2. from Riquelme et al. 2016)

earthquakes, down to Mw 6. Thus, Kanamori and Rivera (Kanamori et al. 2008) used, in addition to their standard frequency band (Kanamori and Rivera 2008), slightly shorter frequency bands, such as 0.00167 Hz (600 s) to 0.005 Hz (200 s) or 0.002 Hz (500 s) to 0.01 Hz (100 s), depending on the noise level of the available data.

Hayes et al. (2009) implemented the W-phase technique at the USGS and extended it to lower magnitudes. Duputel, Z. (Duputel et al. 2011, 2012) calculated W-phase moment tensors for all earthquakes worldwide above 6.5 from 1990 to 2012 and recalculated new filters for different magnitudes. Rivera et al. (2011) and Riquelme et al. (2016) extended the algorithm to regional data from high-rate GPS data (see Fig. 17c). They demonstrated that inverting for the W-phase CMT using high-rate, continuous cGPS data from stations within 5° from the earthquake source would provide a full moment tensor within 5 min of rupture initiation, depending on the number of stations and azimuthal coverage. Regional implementations are also operating in Japan, Mexico, Australia, Taiwan, and in China. PTWC

scientists are currently testing the W-phase inversions from data obtained within 30 degrees epicentral distance, in order to obtain the W-phase CMT and Mw within 15 min of earthquake origin time.

In addition to size, source mechanism, and source duration, the warning centers are interested in more detailed properties of the source such as direction of rupture and the distribution of slip along the fault. This information is important as it can be used in tsunami wave height forecast models to better their predictions.

In the near future, the tsunami warning centers will incorporate finite-fault modeling. Finite-fault modeling involves the inversion of seismic waveforms to recover more detailed information about the source process including the slip distribution, rupture propagation speed, and moment release history (Ammon et al. 2005; Giovanni et al. 2002; Hartzell and Heaton 1986; Hartzell and Mendoza 1991; Mendoza 1996; Wald et al. 1990).

Weinstein and Lundgren (2008) explored the potential of a simple teleseismic P-wave inverse method for the rupture history of an earthquake for use in a tsunami warning center context. The calculations proceed quickly enough that a slip

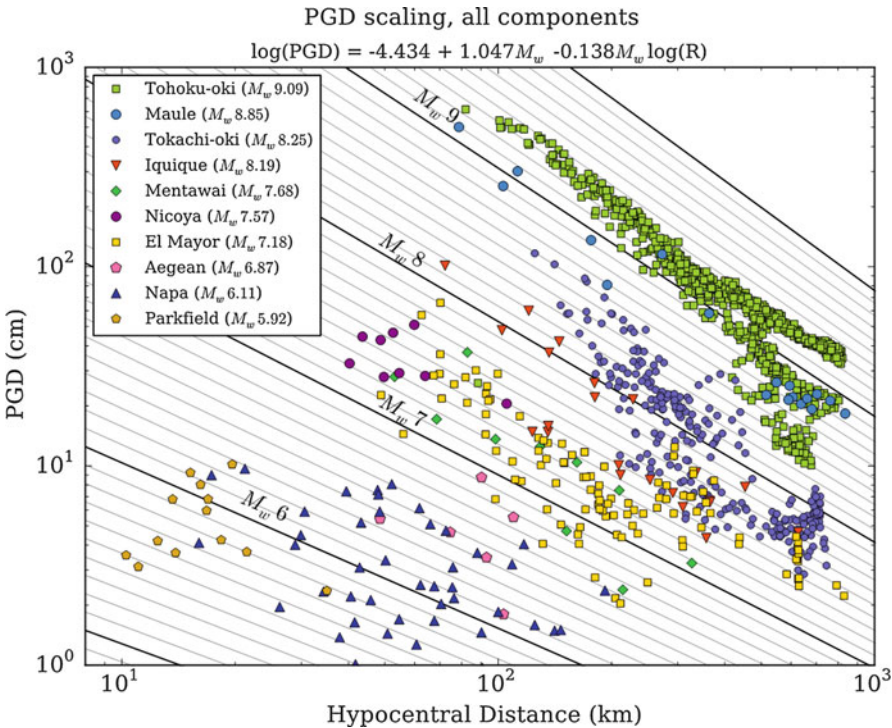
distribution may be available just a few minutes after a suitable set of P-waveforms are obtained. Hence finite-fault modeling results can be used in tsunami wave height forecast models to provide a timely initial estimate of tsunami wave heights.

Using the seismic data available within the first 5 min after rupture initiation, tsunami warning centers may significantly underestimate the magnitude of great Mw 9 plus earthquakes such as the Mw 9.2 Sumatra 2004 event, complex events such as the Mw 8.4 Peru 2001 earthquake, and of “tsunami earthquakes” such as the 1992 Nicaragua and 2006 Java earthquakes, leading to inaccurate tsunami forecasts for those most threatened by the resulting tsunami (Figs. 4, 5, and 6).

Global Navigation Satellite System (GNSS) displacement data and seismogeodetic (Bock et al. 2011) in the near field will enable tsunami warnings within 3–4 min after initiation of rupture at the hypocenter (Melgar et al. 2016). Real-time source modeling codes already provide unsaturated

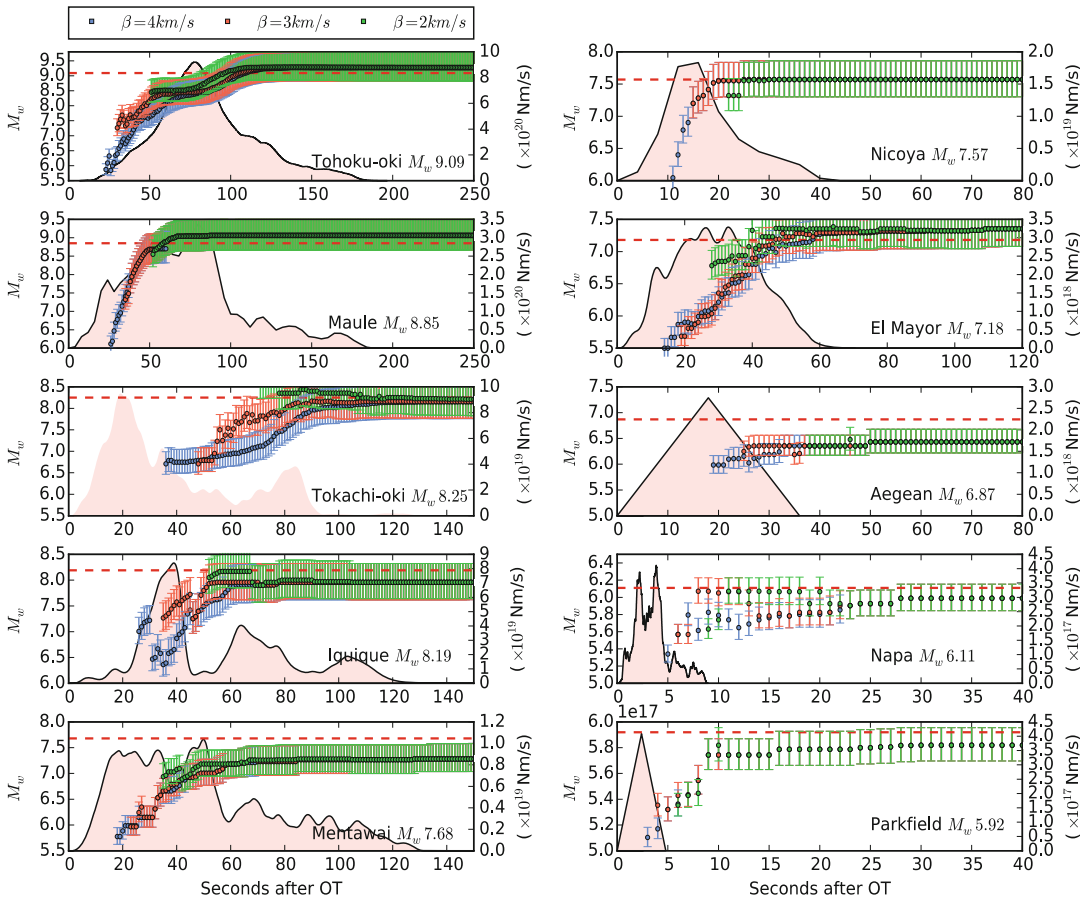
estimates of moment magnitude from peak ground displacement (PGD) (Crowell et al. 2013; Melgar and Crowell 2013; Melgar et al. 2015), centroid moment tensors (Melgar et al. 2012; Melgar and Crowell 2013; Rivera et al. 2011; Riquelme et al. 2016), and fault slip inversions (Crowell et al. 2012; Melgar and Crowell 2013; Melgar et al. 2016), within 1–2 min of rupture initiation (Figs. 18, 19 and 20). This source information can then drive PTWC’s real-time tsunami forecasting model (RIFT) (Duputel et al. 2011; Fryer et al. 2010; Savage 1972; Scholz 1982; Wang et al. 2012), generating forecasts of expected tsunami amplitudes at the near-source coasts within 3–4 min of earthquake origin time (Melgar et al. 2016).

NASA/JPL, the Scripps Institute of Oceanography (SIO), Central Washington University (CWU), University of Oregon, and the University of Washington (UW), with funding and guidance from NASA and leveraging the USGS funded ShakeAlert development, are providing this data



Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings, Fig. 18 Scaling of peak ground displacement (PGD) measurements. The

oblique lines are the predicted scaling values from the L1 regression of the PGD measurements as a function of hypocentral distance. (Figure 2 from Melgar 2015)



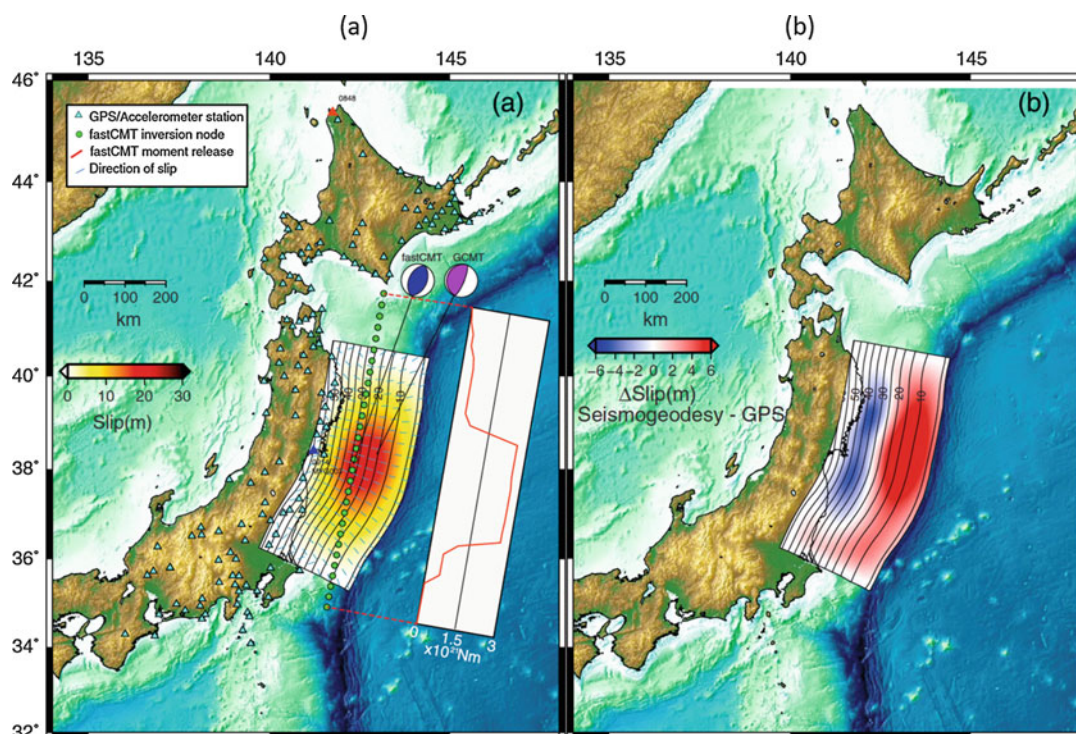
Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings, Fig. 19 Retrospective analysis of the time evolution of magnitude using the scaling law of equation (1) from Melgar et al. (2015). Plotted are the magnitude calculations using three travel time masks at 2, 3, and 4 km/s. The error bars are

determined using the uncertainties of the regression coefficients. The red dashed line is the magnitude from the slip inversion for that event. The shaded pink regions are the source time functions for the kinematic slip inversions. (Figure 4. From Melgar 2015)

and source characterization algorithms to the NOAA National Tsunami Warning Center (NTWC) and to the Pacific Tsunami Warning Center (PTWC).

In the first 2 years of this work, we have established and deployed an Earthworm architecture (GWORM) for this data movement for the algorithms. We have also installed modules for estimating moment magnitude (M_w) from peak ground displacement (PGD), within 2–3 min of the event, computing coseismic displacements converging to static offsets and visualization tools. These modules are being installed in the TWC's operational framework and will be tested using both historical data recordings and new events as they occur.

Global Navigation Satellite System (GNSS) data, and the real-time earthquake source modeling codes which rely on it, will provide unsaturated estimates of moment magnitude, centroid moment tensor solutions, and fault slip models to NOAA's tsunami warning centers within a few minutes after earthquake initiation. These earthquake source characterizations will drive tsunami models such as RIFT (Duputel et al. 2011; Fryer et al. 2010; Savage 1972; Scholz 1982; Wang et al. 2012) (discussed above), enabling tsunami warnings within 3–5 min after earthquake origin time for coastal populations in the near field, where over 80% of tsunami-related casualties would otherwise occur.



Earthquake Source Parameters: Rapid Estimates for Tsunami Forecasts and Warnings, Fig. 20

CMT and slip inversions from seismogeodetic data (GNSS and seismic data combined via Kalman filter). (a) “fastCMT” and slip inversion results from GNSS data only. Green circles are the point sources superimposed to compute the line source of CMT solutions, the final averaged solution shown as fastCMT, and the Global CMT solution (GCMT) shown for comparison. The inset shows the moment release from the line source as a function of distance along fault. Shown along the fault interface with 10 km depth contours from the

Slab 1.0 model (Hayes et al. 2011) is the result of the slip inversion; the blue lines represent the direction of slip. The triangles indicate the locations of all the GPS/accelerometer stations used for computing the slip inversion and CMT solution. (b) Difference between slip inversions computed using seismogeodetic displacements compared to the inversion carried out using the GPS-only derived displacements. Red indicates more slip with the seismogeodetic solution, and blue indicates more slip with the GPS-only solution. (Fig. 2 from Melgar et al. 2013)

Acknowledgments We thank Paula Dunbar of the National Geophysical Data Center for helping us obtain the data we needed and IRIS without which most of this work let alone a functioning TWC would not be possible. We also thank the PTWC for use of its facilities in preparing this manuscript.

Bibliography

Primary Literature

Aki K (1966) Generation and propagation of G waves from the Niigata earthquake of June 16, 1964, Part 2. Estimation of earthquake moment, from the G wave spectrum. *Bull Earthquake Res Inst Tokyo Univ* 44:73–88
 Aki K (1967) Scaling law of seismic spectrum. *J Geophys Res* 72:1217–1231

Allen RV (1978) Automatic earthquake recognition and timing from single traces. *Bull Seism Soc Am* 68:1521–1532
 Allen RV (1982) Automatic phase pickers: their present use and future prospects. *Bull Seism Soc Am* 72:225–242
 Ammon CJ, Ji C, Thio HK, Robinson D, Sidao N, Hjorleifsdottir V, Kanamori H, Lay T, Das S, Helmberger D, Ichinose G, Polet J, Wald D (2005) Rupture process of the 2004 Sumatra–Andaman earthquake. *Science* 6:113–1139
 Arakawa A, Lamb V (1977) Computational design of the basic dynamical processes of the UCLA general circulation model. *Methods Comput Phys* 17:174–267
 Becker JJ, Sandwell DT, Smith WHF, Braud J, Binder B, Depner J, Fabre D, Factor J, Ingalls S, Kim S-H, Ladner R, Marks K, Nelson S, Pharaoh A, Trimmer R, Von Rosenberg J, Wallace G, Weatherall P (2009) Global Bathymetry and Elevation Data at

- 30 Arc Seconds Resolution: SRTM30_PLUS. Marine Geodesy 32(4):355–371. <https://doi.org/10.1080/01490410903297766>
- Bilek SL, Lay T (1999) Rigidity variations with depth along interplate megathrust faults in subduction zones. *Nature* 400:443–446
- Bilek SL, Ruff LJ (2002) Analysis of the 23 June 2001 *M*_W 8.4 Peru underthrusting earthquake and its aftershocks. *Geophys Res Lett* 8:21:1–21:4
- Bilek SL, Lay T, Ruff LJ (2004) Radiated seismic energy and earthquake source duration variations from teleseismic source time functions for shallow subduction zone thrust earthquakes. *J Geophys Res* 109:B09308
- Boatwright J, Choy GL (1986) Teleseismic estimates of the energy radiated by shallow earthquakes. *J Geophys Res* 91:2095–2112
- Boatwright J, Choy GL, Seekins LC (2002) Regional estimates of radiated seismic energy. *Bull Seism Soc Am* 92:1241–1255
- Bock Y, Melgar D, Crowell BW (2011) Real-time strong-motion broadband displacements from collocated GPS and accelerometers. *Bull Seismol Soc Am* 101:2904–2925. <https://doi.org/10.1785/0120110007>
- Bormann P, Baumbach M, Bock G, Grosser H, Choy GL, Boatwright J (2002) Seismic sources and source parameters. In: Bormann P (ed) *IASPEI new manual seismological observatory practice*, vol 1. Chapter 3. *Geoforschungszentrum Potsdam*, pp 1–94. (This can also go as in the text cited review paper)
- Bormann P, Wylegalla K (2005) Quick estimator of the size of great earthquakes. *Eos Trans AGU* 86(46):464
- Bormann P, Wylegalla K, Saul J (2006) Broadband body-wave magnitudes *m*_B and *m*_{BC} for quick reliable estimation of the size of great earthquakes. *USGS Tsunami Sources Work-shop 2006*, poster, http://spring.msi.umn.edu/USGS/Posters/Bormann_et_al_poster.pdf
- Bormann P, Saul J (2008a) Earthquake magnitude. In: Meyers A (ed) *Encyclopedia of complexity and systems science*. Springer, Heidelberg
- Bormann P, Saul J (2008b) The new IASPEI standard broadband magnitude *m*_B. *Seism Res Lett* 79:698–705
- Brune J (1970) Tectonic stress and seismic shear waves from earthquakes. *J Geophys Res* 75:4997–5009
- Brune J (1971) Tectonic stress and seismic shear waves from earthquakes; Correction. *J Geophys Res* 76:5002
- Bryant E (2001) Distribution and fatalities. In: Bryant E (ed) *Tsunami: the underrated hazard*. School of Science and the Environment, Coventry University, Coventry. Cambridge University Press, Cambridge, pp 15–24
- Choy GL, Boatwright J (2007) The energy radiated by the 26 December 2004 Sumatra–Andaman earthquake estimated from 10-minute P-wave windows. *Bull Seism Soc Am* 97:S18–S24
- Crowell BW, Bock Y, Melgar D (2012) Real-time inversion of GPS data for finite fault modeling and rapid hazard assessment. *Geophys Res Lett* 39:L09305. <https://doi.org/10.1029/2012GL051318>
- Crowell BW, Melgar D, Bock Y, Haase JS, Geng J (2013) Earthquake magnitude scaling using seismogeodetic data. *Geophys Res Lett* 40:6089–6094. <https://doi.org/10.1002/2013GL058391>
- Cummins PR (1997) Earthquake near field and W phase observations at teleseismic distances. *Geophys Res Lett* 24:2857–2860
- Duputel Z, Rivera L, Kanamori H, Hayes GP, Hirshorn B, Weinstein S (2011) Real-time W-phase inversions during the 2011 Tohoku-oki earthquake. *Earth Planets Space* 63(7):535–539
- Duputel Z, Rivera L, Kanamori H, Hayes GH (2012) W phase source inversion for moderate to large earthquakes (1990–2010). *Geophys J Int* 189:1125–1147
- Ebeling C, Okal E (2012) An extension of the E/M0 tsunami earthquake discriminant Theta to regional distances. *Geophys J Int* 190(3):1640–1656
- Evans JR, Allen S (1983) A teleseism-specific detection algorithm for single short-period traces. *Bull Seism Soc Am* 73:1173–1186
- Foster JH, Brooks BA, Wang D, Carter GS, Merrifield MA (2012) Improving Tsunami Warning Using Commercial Ships. *Geophys Res Lett* 39:L09603. <https://doi.org/10.1029/2012GL051367>
- Fryer G, Hirshorn B, McCreery S, Cessaro RK, Weinstein S (2005) Tsunami warning in the near field: the approach in Hawaii. *EOS* 86:S44B-04
- Fryer GJ, Holschuh ND, Wang D, Becker N (2010) In: Improving tsunami warning with a rapid linear model, Abstract NH33A-1378 presented at 2010 fall meeting. AGU, San Francisco, 13–17 Dec
- Fukao Y (1979) Tsunami earthquakes and subduction processes near deep-sea trenches. *J Geophys Res* 84:2303–2314
- Geller RJ, Kanamori H (1977) Magnitudes of great shallow earthquakes from 1904 to 1952. *Bull Seismol Soc Am* 67:587–598
- Gica E, Spillane VTMC, Chamberlin C, Newman J (2008) In: Development of the forecast propagation database for NOAA's Short-term Inundation Forecast for Tsunamis (SIFT). NOAA Technical Memorandum OAR PMEL-139, Pacific Marine Environmental Laboratory, Seattle, p 95
- Giovanni MK, Beck SL, Wagner L (2002) The June 23, 2001 Peru earthquake and the southern Peru subduction zone. *Geophys Res Lett* 29:14:1–14:4
- Global Centroid Moment Tensor Catalog, <http://www.globalcmt.org>
- Gower J (2005) Jason 1 detects the 26 December 2004 tsunami. *Eos Trans AGU* 86(4):37–38
- Gutenberg B (1945a) Amplitudes of surface waves and magnitudes of shallow earthquakes. *Bull Seism Soc Am* 35:3–12
- Gutenberg B (1945b) Amplitudes of P, PP, and S, and magnitudes of shallow earthquakes. *Bull Seism Soc Am* 35:57–69
- Gutenberg B (1945c) Magnitude determinations of deep-focus earthquakes. *Bull Seism Soc Am* 35:117–130
- Gutenberg B, Richter CF (1956a) Earthquake magnitude, intensity, energy and acceleration. *Bull Seism Soc Am* 46:105–145

- Gutenberg B, Richter CF (1956b) Magnitude and energy of earthquakes. *Annali di Geofisica* 9:1–15
- Hanks T, Kanamori H (1979) A moment magnitude scale. *J Geophys Res* 84:2348–2350
- Hara T (2007a) Measurement of the duration of high-frequency radiation and its application to determination of the magnitudes of large shallow earthquakes. *Earth Planets Space* 59:227–231
- Hara T (2007b) Magnitude determination using duration of high frequency energy radiation and displacement amplitude: application to tsunami earthquakes. *Earth Planets Space* 59:561–565
- Hartzell S, Heaton T (1986) Rupture history of the 1984 Morgan Hill, California, earthquake from the inversion of strong motion records. *Bull Seism Soc Am* 76:649–674
- Hartzell S, Mendoza C (1991) Application of an iterative least-squares waveform inversion of strong-motion and teleseismic records to the 1978 Tabas, Iran earthquake. *Bull Seism Soc Am* 81:305–331
- Hayes GP, Rivera L, Kanamori H (2009) Source inversion of the W-Phase: real-time implementation and extension to low magnitudes. *Seismol Res Lett* 80(5):817–822
- Hayes GP, Earle PS, Benz HM, Wald DJ, Briggs R, The USGS/NEIC Earthquake Response Team (2011) 88 hours: the U.S. Geological survey national earthquake information center response to the March 11, 2011 Mw 9.0 Tohoku earthquake. *Seismol Res Lett* 82(4):481–493
- Henry C, Das S (2001) Aftershock zones of large shallow earthquakes: fault dimensions, aftershock area expansion and scaling relations. *Geophys J Int* 147:272–293
- Hirshorn B, Lindh A, Allen R (1987) Real time signal duration magnitudes from low-gain short period seismometers. *USGS OFR* 87:630
- Hirshorn B, Lindh G, Allen RV, Johnson C (1993) Real time magnitude estimation for a prototype early warning system (EWS) from the P-wave, and for earthquake hazards monitoring from the coda envelope. *Seis Res Lett* 64:48
- Hirshorn B, Weinstein S, Tsuboi, S. (2013). On the application of Mwp in the near field and the March 11, 2011 Tohoku earthquake. *Pure Appl Geophys* 170 (6–8):975–991. <https://doi.org/10.1007/s00024-012-0495-3>
- Hirshorn B (2004) Moment magnitudes from the initial P-wave for local tsunami warnings. *Seism Res Lett* 74:272–273
- Hirshorn B (2007) The Pacific tsunami warning center response to the Mw6.7 Kiholo Bay earthquake and lessons for the future. *Seism Res Lett* 78:299
- Hoechner A, Ge M, Babeyko Y, Sobolev SV (2013) Instant tsunami early warning based on real-time GPS – tohoku 2011 case study. *Nat Hazards Earth Syst Sci* 13:1285–1292
- Houston H, Kanamori H (1986) Source spectra of great earthquakes, teleseismic constraints on rupture process and strong motion. *Bull Seism Soc Am* 76:19–42
- Imamura F (1996) Review of tsunami simulation with a finite difference method. In: Yeh H, Liu P, Synolakis C (eds) Long-wave runup models. World Scientific, Singapore, pp 25–42
- Ishii M, Shearer PM, Houston H, Vidale JE (2005) Extent, duration and speed of the 2004 Sumatra–Andaman earthquake imaged by the Hi-Net array. *Nature* 435:933–936
- Johnson CE, Lindh A, Hirshorn B (1994) Robust regional phase association. *US Geol Surv Open-File Rept* 94:621
- Johnson CE, Bittenbinder A, Bogaert B, Dietz L, Kohler W (1995) Earthworm: a flexible approach to seismic network processing. *IRIS Newslett XIV* 2:1–4
- Kamigaichi O (2009) Tsunami forecasting and warning. In: Meyers RA (ed) *Encyclopedia of Complexity and Systems Science*. Springer, Heidelberg, pp 9592–9618
- Kanamori H (1972) Mechanism of tsunami earthquakes. *Phys Earth Planet Inter* 6:246–259
- Kanamori H (1977) The energy release in great earthquakes. *J Geophys Res* 82:2981–2987
- Kanamori H (1978) Quantification of earthquakes. *Nature* 271:411–414
- Kanamori H (1983) Magnitude scale and quantification of earthquakes. *Tectonophysics* 93:185–199
- Kanamori H (1993) W Phase. *Geophys Res Lett* 20:1691–1694
- Kanamori H, Kikuchi M (1993) The 1992 Nicaragua earthquake: a slow tsunami earthquake associated with subducted sediment. *Nature* 361:714–715
- Kanamori H (2006) Seismological aspects of the December 2004 great Sumatra Andaman earthquake. *Earthquake Spectra* 22:S1–S12
- Kanamori H, Rivera L (2008) Source inversion of W phase – speeding up tsunami warning. *Geophys J Int* 175:222–238
- Kanamori H, Rivera L, Earthquake Research Institute (2008) Application of the W phase source inversion method to regional tsunami 1 Seismological Lab, California Institute of Technology, Pasadena, University of Tokyo
- Kikuchi M, Yamanaka Y (2001) EIC Seismological Note Number 105, www.eic.eri-u-tokyo.ac.jp/EIC/EIC_news/105E.html
- Krüger F, Ohrnberger M (2005) Tracking the rupture of the Mw9.3 Sumatra earthquake over 1150 km at teleseismic distance. *Nature* 435:937–939
- Kowalik Z, Whitmore PM (1991) An investigation of two tsunamis recorded at Adak, Alaska. *Sci Tsunami Haz* 9:67–84
- Lamb H (1932) *Hydrodynamics*, 6th edn. Dover Publications, New York
- Lay T, Kanamori H, Ammon C et al (2005) The Great Sumatra – Andaman earthquake of 26 December 2004. *Science* 308:1127–1133
- Lockwood OG, Kanamori H (2006) Wavelet analysis of the seismograms of the 2004 Sumatra–Andaman earthquake and its application to tsunami early warning. *Geochem Geophys Geosyst* 7:Q09013. <https://doi.org/10.1029/2006GC001272>
- Lomax A, Michelini A, Piatanesi A (2007) An energy-duration procedure for rapid determination of

- earthquake magnitude and tsunamigenic potential. *Geophys J Int* 170:1195–1120
- Lomax A, Michelini A (2008) Mwpd: A duration-amplitude procedure for rapid determination of earthquake magnitude and tsunamigenic potential from P waveforms. *Geophys J Int*. <https://doi.org/10.1111/j.1365-246X.2008.03974>
- Lomax A, Michelini A (2012) Tsunami early warning within 5–10 minutes. *Pure Appl Geophys* 169. <https://doi.org/10.1007/s00024-012-0512-6>
- Melgar D, Bock Y, Crowell BW (2012) Real-time centroid moment tensor determination for large earthquakes from local and regional displacement records. *Geophys J Int* 188:703–718. <https://doi.org/10.1111/j.1365-246X.2011.05297.x>
- Melgar D, Crowell BW, Bock Y, Haase JS (2013) Rapid modeling of the 2011 Mw 9.0 Tohoku-oki earthquake with seismogeodesy. *Geophys Res Lett* 40:1–6
- Melgar D, Crowell BW, Geng J, Allen RM, Bock Y, Riquelme S, Hill EM, Protti M, Ganas A (2015) Earthquake magnitude calculation without saturation from the scaling of peak ground displacement. *Geophys Res Lett* 42:5197–5205. <https://doi.org/10.1002/2015GL064278>
- Melgar D et al (2016) Local tsunami warnings: perspectives from recent large events. *Geophys Res Lett* 43:1109–1117. <https://doi.org/10.1002/2015GL067100>
- Mendoza C (1996) Rapid derivation of rupture history for large earthquakes. *Seismol Res Lett* 67:19–26
- NOAA Special Report (2012) Proceedings and results of the 2011 NTHMP model benchmarking workshop, Department of Commerce and Texas A&M University at Galveston, p 437
- Newman AV, Okal EA (1998) Teleseismic estimates of radiated seismic energy: the E/M0 discriminant for tsunami earthquakes. *J Geophys Res* 103:26885–26898
- Okada Y (1985) Surface deformation due to shear and tensile faults in a half space. *Bull Seismol Soc Am* 75:1135–1154
- Okal EA, Talandier J (1989) Mm: a variable-period mantle magnitude. *J Geophys Res* 94:4169–4193
- Okal EA (1992a) Use of mantle magnitude Mm for reassessment of the moment of historical earthquakes-I: Shallow events. *PAGEOPH* 139:17–57
- Park J, Song TA, Tromp J, Okal E, Stein S, Rault G, Clevede E, Laske G, Kanamori H, Davis P, Berger J, Braitenberg C, Camp MV, Xiang'e L, Heping S, Houze X, Rosat S (2005) Earth's free oscillations excited by the 26 December 2004 Sumatra–Andaman earthquake. *Science* 20:1139–1144
- Richter CF (1935) An instrumental earthquake magnitude scale. *Bull Seism Soc Am* 25:1–32
- Rivera LA, Kanamori H, Duputel Z (2011) W phase source inversion using the high-rate regional GPS data of the 2011 Tohoku oki earthquake, in AGU fall meeting abstracts, vol 1, p 4
- Riquelme S, Bravo F, Melgar D, Benavente R, Geng J, Barrientos S, Campos J (2016) W phase source inversion using high-rate regional GPS data for large earthquakes. *Geophys Res Lett* 43(7):3178–3185
- Satake K (1994) Mechanism of the 1992 Nicaragua tsunami earthquake. *Geophys Res Lett* 21:2519–2522
- Savage JC (1972) Relation of corner frequency to fault dimensions. *J Geophys Res* 77:3788–3795
- Scholz C (1982) Scaling laws for large earthquakes: consequences for physical models. *Bull Seism Soc Am* 72:1–14
- Seno T, Hirata K (2007) Did the 2004 Sumatra–Andaman earthquake involve a component of tsunami earthquakes? *Bull Seism Soc Am* 97:S296–S306
- Stein S, Okal EA (2007a) Ultralong period seismic study of the December 2004 Indian Ocean earthquake and implications for regional tectonics and the subduction process. *Bull Seism Soc Am* 97:279–295
- Stein S, Okal EA (2007b) Ultralong Period Seismic Study of the December 2004 Indian Ocean earthquake and implications for regional tectonics and the subduction process. *Bull Seism Soc Am* 97:S279–S295
- Tatehata H (1997) The new tsunami warning system of the Japan meteorological agency. In: Hebenstreit G (ed) Perspectives on tsunami hazard reduction. Kluwer, Dordrecht, pp 175–188
- Tsuboi SK, Abe K, Takano K, Yamanaka Y (1995) Rapid determination of Mw from broadband P waveforms. *Bull Seism Soc Am* 83:606–613
- Tsuboi S, Whitmore PM, Sokolowski TJ (1999) Application of Mwp to deep and teleseismic earthquakes. *Bull Seism Soc Am* 89:1345–1351
- Vanek J, Zatopek A, Kamik V, Kondorskaya N, Riznichenko Y, Savarenski S, Solovov S, Shebalin N (1962) Standardization of magnitude scales. *Izv Acad Sci USSR Geophys Ser*:108–111. English translation
- Vassiliou MS, Kanamori H (1982) The energy release in earthquakes. *Bull Seism Soc Am* 72:371–387
- Wald DJ, Helmberger DV, Hartzell S (1990) Rupture process of the 1987 Superstition Hills earthquake from the inversion of strong-motion data. *Bull Seism Soc Am* 80:1079–1098
- Wang D, Walsh D, Becker N, Fryer G (2009) A methodology for tsunami wave propagation forecast in real time, EOS Trans AGU, 90(52), Fall Meet. Suppl., Abstract OS43A-1367
- Wang D, Becker NC, Walsh D, Fryer G, Weinstein S, McCreery C, Sardina V, Hsu V, Hirshorn B, Hayes G, Duputel Z, Rivera L, Kanamori H, Koyanagi K, Shiro B (2012) Real-time forecasting of the April 11, 2012 Sumatra Tsunami. *Geophys Res Lett* 39:L19601. <https://doi.org/10.1029/2012GL053081>
- Weinstein SA, McCreery C, Hirshorn B, Whitmore P (2005) Comment on “A strategy to rapidly determine the magnitude of great earthquakes” by Menke W, Levin V. *Eos* 86:263
- Weinstein SA, Okal EA (2005) The mantle magnitude Mm and the slowness parameter 8: five years of real-time use in the context of tsunami warning. *Bull Seism Soc Am* 95:779–799
- Weinstein SA, Lundgren PL (2008) Finite fault modeling in a tsunami warning center context. In: Tiampo KF,

- Weatherley DK, Weinstein SA (eds) *Earthquakes: simulations, sources and tsunamis*. Birkhauser, Basel
- Wells DL, Coppersmith KJ (1994) New empirical relationships among magnitude, rupture length, rupture width, rupture area, and surface displacement. *Bull Seismol Soc Am* 84:974–1002
- Wessel P (2009) Analysis of observed and predicted tsunami travel times for the Pacific and Indian Oceans. *Pure Appl Geophys* 166:301–324. <https://doi.org/10.1007/s00024-008-0437-2>
- Wessel P, Smith WHF (1991) Free software helps map and display data. *Eos Trans AGU* 72:441
- Whitmore PM, Sokolowski TJ (2002) Automatic earthquake processing developments at the US West Coast/Alaska tsunami warning center. *Recent Research Developments in Seismology*, pp 1–13, Kervala: Transworld Research Net- work. ISBN 81-7895, 072-3
- Whitmore PM, Tsuboi S, Hirshorn B, Sokolowski TJ (2002) Magnitude-dependent correction for Mwp. *Sci Tsunami Hazards* J20:187–192
- Widjo K et al (2006) Rapid survey on tsunami Java July 17 2006. http://nctr.pmel.noaa.gov/java20060717/tsunami-java170706_e.pdf. Accessed July 2008
- Withers M, Aster R, Young C, Beiriger J, Harris M, Trujillo J (1998) A comparison of select trigger algorithms for automated global seismic phase and event detection. *Bull Seism Soc Am* 88:95–106
- Yasuda T, Mase H (2013) Real-time tsunami prediction by inversion method using offshore observed GPS buoy data: Nankaido. *J Waterway Port Coastal Ocean Eng* 139(3):221–231
- Aki K, Richards PG (1980) *Quantitative seismology theory and methods*, vol 2. W.H. Freeman Co, San Francisco
- Aki K, Richards PG (2002) *Quantitative seismology*, 2nd edn. University Science Books, Sausalito
- Båth M (1981) Earthquake magnitude – recent research and current trends. *Earth Sci Rev* 17:315–398
- Bormann P (ed) (2002) *IASPEI new manual of seismological observatory practice*, vol 1 and 2. GeoForschungsZentrum, Potsdam, p 1250
- Duda S, Aki K (eds) (1983) Quantification of earthquakes. *Tectonophysics* 93. Special issue 3(4):183–356
- Kanamori H, Anderson DL (1975) Theoretical basis of some empirical relations in seismology. *Bull Seismol Soc Am* 65(5):1073–1095
- Kanamori H (1994) Mechanics of earthquakes. *Annu Rev Earth Planet Sci* 22:307–237
- Kanamori H, Rivera L (2006) Energy partitioning during an earth-quake. In: Abercrombie R, McGarr A, Kanamori H, Di Toro G (eds) *Earthquakes: radiated energy and the physics of faulting*. Geophysical monograph, vol 170. American Geophysical Union, Washington, DC, pp 3–13
- Kanamori H, Brodsky E (2004) The Physics of earthquakes. *Rep Prog Phys* 67:1429–1496
- Kanamori H (2006) The diversity of the physics of earthquakes. *Proc Japan Acad Ser B* 80:297–316
- Lay T, Wallace TC (1995) *Modern global seismology*. Academic, San Diego
- Okal EA (1992b) A student's guide to teleseismic body wave amplitudes. *Seism Res Lett* 63(N2):169–180
- Richter CF (1958) *Elementary seismology*. WH Freeman Co, San Francisco
- Stein S, Wysession M (2003) *An introduction to seismology, earthquakes, and earth structure*. Blackwell Publishing, Oxford

Books and Reviews

- Abercrombie R, McGarr A, Di Toro G, Kanamori H (eds) (2006) *Earth- quakes: radiated energy and the physics of faulting*. In: *Geophysical monographs* 170. American Geophysical Union, Washington, DC



Tsunami Forecasting and Warning

Osamu Kamigaichi
Japan Meteorological Agency, Tokyo, Japan

Article Outline

Glossary
Definition of the Subject
Introduction
Complexity Problem in Tsunami Forecasting
Components of a Tsunami Early Warning System (TEWS)
Tsunami Early Warning System in Japan
Future Outlook
Bibliography

Glossary

Centroid Moment Tensor solution One of the representation of seismic source process. It is represented by the moment tensor, which is a combination of six independent equivalent force couples, and is a weighted average of the source process in time and space. Since it represents an overall image of the source process, it is suitable to evaluate tsunamigenic potential of the earthquake.

Earthquake early warning Earthquake early warning is to enable countermeasures in advance for strong motion disaster by detecting seismic P wave at stations near the epicenter, quickly estimate seismic intensity and arrival time of S wave, and transmit these estimations before the S-wave arrival. The Japan Meteorological Agency (JMA) started to provide EEW to the general public in October 2007. This technique is applicable to quicken tsunami warning dissemination.

Forecast point In this article, it is defined as the location of the offshore point where tsunami

amplitude is evaluated by using numerical tsunami propagation simulation.

Intergovernmental Coordination Group The group established under UNESCO/IOC to facilitate international cooperation for the tsunami disaster mitigation. There are four ICGs as of now (Pacific Ocean, Indian Ocean, Caribbean Sea, Northeastern Atlantic Ocean, and Mediterranean Sea). One of the most important characteristics of tsunami is that it can cause huge disaster even after long-distance propagation due to amplification near the coast. Therefore, international cooperation, especially the prompt data and information exchange, is essential for the disaster mitigation.

Simulation point In this article, it is defined as the surface projection location of the hypothetical earthquake fault center. The vertical component of ocean bottom deformation due to earthquake fault dislocation calculated by elastic theory gives the initial tsunami waveform for the numerical tsunami propagation simulation.

Tsunami amplitude Amplitude is measured from undisturbed sea level to peak or trough of the wave. By definition, it can be positive or negative. Tsunami amplitude can be measured in real time by instruments like tide gauge, pressure sensor, etc. and can be reproduced by numerical tsunami propagation simulation. One need not be confused with the term “run-up height.” It is the maximum height of inundation on land and measured in post-tsunami field surveys from traces of tsunami (i.e., damage of constructions, vegetative markers, etc.).

Tsunami Early Warning System It consists of four components, namely, (1) seismic network, (2) seismic data processing system, (3) tsunami forecast system, and (4) sea level data monitoring system. In a broader sense, warning transmission system (downlink to disaster management organizations and public) is also included.

Definition of the Subject

Tsunami is, along with strong motion, one of the two major disasters caused by earthquake. To mitigate tsunami disaster, it is important to integrate software countermeasures like tsunami forecast to enable timely evacuation from area at risk before tsunami strikes the coast, as well as to intensify hardware countermeasures particularly in vulnerable coastal areas like building banks and water gates. Tsunami disaster mitigation can be achieved effectively by the appropriate combination of the software and hardware countermeasures. Also, improving people's awareness on the tsunami disaster, necessity of spontaneous evacuation when they notice an imminent threat of tsunami on their own (feeling strong shaking near the coast, seeing abnormal sea level change, etc) and how to respond to the tsunami forecast, and conducting tsunami evacuation drill are very important issues for disaster mitigation.

In this article, the tsunami forecast, as the most typical software countermeasure that national organization can provide, is mainly described. Recent progress in science deepened our understanding of the tsunami-generating source mechanisms, tsunami propagation, and inundation process and enabled the development of sophisticated numerical simulation programs. But at the same time, complexities of focal process and tsunami behavior, especially near the coast, have been revealed also. Therefore, careful consideration of the complicated nature of tsunami phenomena is essential in order to make tsunami forecast contents and the dissemination of warnings effective for disaster mitigation.

Introduction

Tsunami is an oceanic gravity wave generated by submarine fault dislocation or other origins such as mud or rock slumps on steep continental margin slopes, marine volcanic eruptions, and others. A large tsunami may cause disasters along densely populated or built-up coasts and sometimes also by the inundation of low land areas, up to several km inland. Most tsunamis are

generated by large earthquakes occurring in oceanic areas, and it is possible to estimate tsunami generation to a certain extent from seismic wave analysis. By taking advantage of the propagation velocity difference between the much faster seismic and the slower tsunami waves, it is possible to mitigate tsunami disasters by issuing tsunami forecast before the tsunami arrives at the coast, thus enabling evacuation and other countermeasures. For other origin of tsunami, it is still difficult to quantitatively forecast tsunami generation until it is observed actually by the sea level change sensors.

In earlier years, an empirical method had been used to estimate tsunami amplitude via a regression formula that relates tsunami amplitude to the magnitude of the earthquake and its epicentral distance from the coast of interest. Recently, significant progress has been made in the understanding of the tsunami characteristics. This permits numerical simulation of the tsunami propagation once the initial tsunami wave distribution in the source area is correctly known, and bathymetry data are available with sufficient spatial resolution. In Japan, numerical simulation technique has been used in the operational tsunami forecasting procedure in the Japan Meteorological Agency (JMA) since 1999. Efforts have been made also in other countries to incorporate numerical simulation technique in tsunami forecasting. In this article, mainly based on JMA experience, procedures are introduced how to conduct tsunami forecast service by using numerical simulation techniques and giving due consideration to the complexity of this problem.

Complexity Problem in Tsunami Forecasting

The complexity of tsunami forecasting with numerical methods is twofold and mainly due to the following:

- The uncertainty of the initial tsunami wave distribution in the source area
- The complexity of bathymetry and coastal topography

Uncertainty of the Initial Tsunami Wave Distribution

For local tsunami, only limited time is left from the generation of tsunami to its arrival at the coast. In order to assure maximum lead time for evacuation, this necessitates the tsunami forecast to be based on the earliest available data from seismic wave analysis. On the other hand, in the case of a distant tsunami sources, a relatively long lead time is left until the tsunami strikes the coast, and data of sea level change can be recorded in near real time by tidal stations on the way from the source to the coast. This allows the tsunami forecast to be based on the actual observation of generated tsunami waves and thus to reduce the uncertainty in the initial spatial distribution of the tsunami wave in the source area. Also for local events, it is possible to improve the accuracy and reliability of the tsunami forecast in a step-by-step manner, if more detailed data and reliable analysis results become available after the first forecast has been issued. Therefore, the most practical approach in view of the disaster mitigation is to assure the rapid issuance of the first forecast based on the seismic wave analysis and then to update it with improved data.

There are several uncertainties when the initial tsunami wave distribution is inferred from seismological data analysis alone.

Uncertainty of the Relative Location of the Hypocenter in the Rupture Area

The hypocenter of an earthquake is merely the location from where the rupture starts. It is calculated from first arriving seismic waves. Depending on the direction of rupture propagation (e.g., unilateral, bilateral, radial) and the complexity/irregularity of the actual fault rupture, the seismologically determined hypocenter may neither lie in the center of the rupture area nor coincide with the area of maximum fault displacement. The bigger the earthquake and thus the larger the spatial extent of the rupture area, the less representative is the seismologically derived hypocenter as a parameter for characterizing the earthquake rupture is. To reduce this uncertainty, some methods have been developed, like a rupture process inversion by using teleseismic broadband seismogram (Kikuchi and

Kanamori 1991) or near-field strong motion data (Ide et al. 1996). Fault model inversion by using a coseismic step estimated by GPS data analysis has been developed (Matsu'ura and Hasegawa 1987; Ozawa 1996). And the spatial extent of the tsunami source area can also be constrained by using inverse refraction diagrams from coastal tidal stations (Abe 1973), ocean bottom pressure sensors (DART (Deep-ocean Assessment and Reporting of Tsunamis) type (González et al. 2005) and cable type (Kanazawa 2013; Kaneda 2010)), or GPS tsunami meters (Kawai et al. 2013) that recorded tsunami arrivals as well as by satellite sea surface altimetry data taken over the source area (Jim Gower 2005; Hayashi 2008). By utilizing offshore tsunami meter data which is less affected by complicated bathymetry in a shallow sea, tsunami source area can be constrained more accurately (Hayashi et al. 2011).

In general, the rupture process is complicated, and the slip distribution on the fault is not uniform. Nonuniformity of the slip distribution can make the initial tsunami wave distribution much more complicated, especially in its short-wavelength component, and may strongly affect the tsunami behavior on the coast. To estimate the slip distribution, in addition to the seismological and geodetic methods mentioned above, fault slip distribution inversion techniques have been developed which compare actually observed tsunami waveforms with pre-calculated theoretical tsunami waveforms from unit fault dislocation segments as Green functions (Satake 1989) (► [“Tsunamis, Inverse Problem of”](#)).

Uncertainty of the Magnitude

The maximum amplitude of seismic body or surface-wave group is usually used for estimating the earthquake magnitude (see this volume) as a measure of earthquake size in a certain period range, which is often limited by the bandwidth of the seismometer response. Therefore, the tsunamigenic potential may be underestimated, especially for gigantic earthquakes (such as the Sumatra-Andaman Mw9.3 Earthquake of December 26, 2004) and “tsunami earthquakes” that generate bigger tsunami than expected from its shaking potential and magnitude values in the short-period range. To overcome this difficulty,

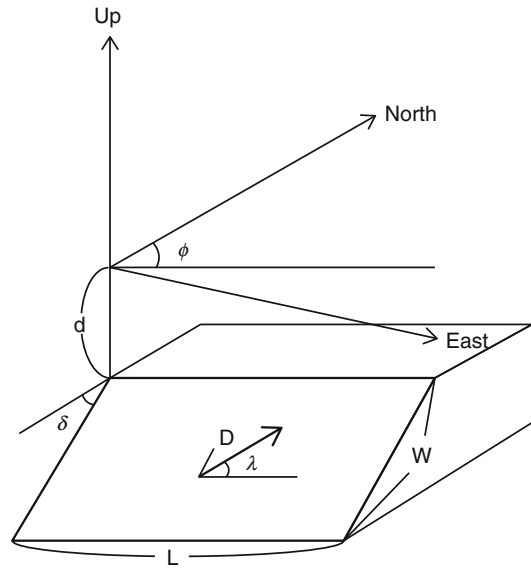
some fast body-wave methods have been developed that consider besides single maximum amplitudes also multiple rupture amplitudes over the rupture duration (Bormann and Wylegalla 2005), earthquake magnitude, integrated seismic energy, and/or total rupture duration time (Hara 2007; Lomax et al. 2007), while others use successfully long-period spectral mantle surface-wave amplitudes in real-time applications in the context of tsunami warning (Weinstein and Okal 2005) and W phase for quick estimation of CMT solution for a practical tsunami warning service (Kanamori and Rivera 2008). Besides these seismological methods for estimating earthquake magnitude, geodetic methods that use the coseismic step or ultra-long-period component observed by crustal deformation sensors (Japan Meteorological Agency 2005; Kasahara and Sasatani 1986) play an increasing role for estimating the size of great earthquakes.

Uncertainty of the Fault Parameters

It is difficult to estimate fault geometry promptly after the earthquake occurrence. Sea floor deformation due to fault motion is usually assumed to be identical to initial tsunami wave distribution. Accordingly, the latter is significantly affected by the specifics of fault geometry, fault and slip orientation in space, as well as rupture complexity. These complexities are usually neglected, and the real fault rupture is roughly approximated by a dislocation model that is represented by a rectangular fault that is described by six parameters, namely, the fault length L , width W , average slip amount D , and the three angles of fault dip δ , strike Φ , and slip (rake) λ (Fig. 1).

Presently, efforts have been made to estimate fault parameters promptly after the earthquake occurrence by taking advantage of evolving high sampling rate (1Hz) real-time GNSS/GPS data processing technology and by detecting coseismic displacements at stations surrounding focal area (Nishimura et al. 2010; Ohta et al. 2012; Yahagi et al. 2014).

Fault Length, Width, and Average Slip Amount Different scaling laws exist



Tsunami Forecasting and Warning, Fig. 1 Definitions of fault parameters. L , W , D , and d denote the fault length, width, average slip amount, and depth of the *top* margin of the fault, respectively. The fault dip δ is measured down from horizontal, strike Φ clockwise round from north and slip (rake) λ anti-clockwise round from strike direction along the fault plane (motion of the hanging wall relative to the footwall)

between L , W , and D with earthquake magnitude (e.g., Bormann et al. 2002; Geller 1976; Kanamori and Anderson 1975; Utsu et al. 2001; Wells and Coppersmith 1994). They allow us to roughly estimate these values from determined magnitudes. Such scaling relations are based on the assumption of more or less constant stress drop (i.e., high stress drop earthquake has larger D for the same L and W of low stress drop earthquake). But the stress drop is different between inter-plate thrust events and intra-plate events. In general, intra-plate event has higher stress drop than inter-plate event. Such differences may cause significant errors when estimating these fault parameters via scaling relations with magnitudes. Best estimates of these parameters, which characterize the earthquake size, are possible via the determination of the seismic moment $M_0 = \mu D A$ with μ being the rigidity of the crustal/lithosphere material in the rupture area $A = L \times W$, and L and W estimation by rupture process analysis described in subsection

“Uncertainty of the Relative Location of the Hypocenter in the Rupture Area.”

Fault Dip, Strike, and Slip Angle Representative values for dip, strike, and slip angles can be set based on the analysis of past great earthquakes having occurred in each region. The fault plane of a newly occurred earthquake is likely to be close to either one of the two nodal planes derived from Centroid Moment Tensor (CMT) solutions. But to select from these two candidates the real acting fault plane, precise aftershock location or rupture process analysis as described in subsection “Uncertainty of the Relative Location of the Hypocenter in the Rupture Area” is necessary.

Complexity Due to Complicated Bathymetry

If the spatial distribution of the initial tsunami wave is given correctly, the tsunami propagation can be forecast in a deterministic manner by using numerical simulation technique. In a long-wave approximation, which considers only wavelengths of the tsunami that are much larger than the sea depth, the propagation velocity v of the tsunami is represented by $v = \sqrt{gd}$, where $g = 9.81 \dots \text{m/s}^2$ is the Earth's gravity acceleration and d the sea depth in m. This approximation is valid for most tsunami cases. As wavelength is the product of velocity times period, tsunami wavelength is proportional to the square root of sea depth. Accordingly, the wavelength of the tsunami become shorter as it approaches the coast because the sea depth becomes shallower, and the wave becomes much more sensitive to finer bathymetry changes and coastal feature. Therefore, a finer mesh of bathymetry and coastal features is necessary near the coast in order to represent the tsunami behavior correctly. But, even if very fine-mesh bathymetry data are available, it is not appropriate to use it in an early stage of tsunami propagation simulation when the initial tsunami wave distribution is still uncertain as mentioned in subsection “Uncertainty of the Initial Tsunami Wave Distribution.”

Further, very fine-mesh simulations require substantially long time for the conduct of simulation. Time constraints are crucial when

incorporating numerical simulation techniques in an operational tsunami forecast procedure, except for distant events. Therefore, especially for local events, it is most practical to conduct tsunami simulations for a variety of scenarios in advance, to store these results in a database, and to conduct tsunami forecast by retrieving the most appropriate case for the determined hypocenter. As described later, the JMA has adopted this way. Even in that case, the mesh size to be used in the simulations must be carefully examined taking into account the required accuracy and spatial resolution in the tsunami forecast for guiding disaster mitigation efforts and the need to complete the simulations for all scenarios in a realistic time span.

When very fine mesh is used, depending on the complexity of the bathymetry and topography near the coast, tsunami amplitude distribution along the coast shows significant scatter, and sometimes, extremely large tsunami amplitude will be estimated very locally. Therefore, it must be clearly defined in advance as to what kind of statistical average value, using a finite number of estimated tsunami amplitudes for the area of interest, should be adopted in the tsunami forecast.

Components of a Tsunami Early Warning System (TEWS)

In general, a Tsunami Early Warning System consists of the following constituents:

1. Seismic network (seismometers and real-time data transmission link)
2. Real-time seismic data processing system for hypocenter and magnitude determination
3. Tsunami forecast system (including warning criteria, assembling of the text, and dissemination)
4. Sea level data monitoring system (tide gauge/tsunami meter and real-time/near-real-time data transmission link)

Such a composition is adopted in Japan, the USA, Russia, Chile, Australia, French

Polynesia, New Caledonia, and in other countries around the Pacific Ocean where the frequency of tsunamigenic earthquake occurrence is the highest among the oceans, and efforts have been made to establish tsunami early warning systems since the middle of twentieth century. The up-to-date status of the TEWS of these countries can be consulted on the UNESCO/Intergovernmental Oceanographic Commission's (IOC) website as based on the national reports submitted to the latest meeting of the Intergovernmental Coordination Group/Pacific Tsunami Warning System (ICG/PTWS). Also in the Indian Ocean countries, after the tremendous tsunami disaster brought by the Great Sumatra Earthquake in 2004, efforts have been made to establish Tsunami Early Warning Systems of the similar composition in different countries. Similar efforts are under way in the Caribbean, Northeastern Atlantic, and Mediterranean regions too, inspired by the Sumatra event and taking into account potential tsunami risk assessments for these areas based on historical records.

To accomplish tsunami warning for local events, local seismic networks (see subsection "[Seismic Network](#)") and real-time seismic data processing systems (see subsection "[Real-Time Seismic Data Processing System](#)") are indispensable. For a distant event, the authorized domestic tsunami warning organization in each country can utilize the hypocenter parameters contained in the international tsunami watch information (e.g., those provided by the Pacific Tsunami Warning Center (PTWC), National Tsunami Warning Center (NTWC) of the USA, and Northwest Pacific Tsunami Advisory Center (NWPTAC) of the JMA in the Pacific and those by INCOIS of India, BMKG of Indonesia, and Joint Australian Tsunami Warning Center (BoM and GA) of Australia in the Indian Ocean).

As a typical example, technical details of Japan's Tsunami Early Warning System, in which all four constituents are fully deployed, are explained in the following section.

As for item 4 (sea level data monitoring system), the USA has been deploying DART

type of buoy (González et al. 2005) system. This is to measure a sea surface vertical displacement by observing the related pressure change at the ocean bottom. DART buoys are placed in far offshore regions where bathymetry is simple, and very simple tsunami waveform can be observed without the influence of complicated bathymetry near a coast. Such measured data are preferable for a comparison with simulated waveforms. And by placing such buoys with proper spacing at a certain distance from the tsunamigenic zone over the deep-ocean basin, tsunami waves, which are sufficiently separated in time from seismic waves, can be observed early enough after the earthquake occurrence to be useful for tsunami EW from distant sources.

The Pacific Marine Environmental Laboratory (PMEL) of the National Oceanic and Atmospheric Administration (NOAA) has been developing a tsunami forecasting system named SIFT (Short-term Inundation Forecasting for Tsunamis) (Titov et al. 2005) based on the DART buoy data. Tsunami simulations are conducted for many different unit faults located along the tsunamigenic zone, and the calculated tsunami waveforms at DART buoy locations are stored in a database as Green functions, together with simulated waveforms at near-shore points. In the case that a tsunami wave is observed by DART buoys, the observed waveforms are represented by a linear combination of several Green functions. Then, tsunami amplitudes at coastal points can be estimated from the linear combination of coefficients and the simulated waveforms at near-shore points followed by nonlinear tsunami inundation computation from points to coasts (see webpage of NOAA/PMEL (<http://nctr.pmel.noaa.gov/>) for applications of their method to recent actual tsunami events).

In Japan, development of similar algorithm focusing on near-field tsunamis, named tsunami Forecasting based on Inversion for initial sea-Surface Height (tFISH), is in progress at the Meteorological Research Institute of JMA (Tsushima et al. 2009). They conduct tsunami simulations originating from unit vertical sea

surface displacements in a unit sea area, and the resulting simulated waveforms at offshore tsunami meters and coastal points of interest are stored as Green functions in a database. They do not adhere to a specific fault model, and, consequently, this method can be applied to nonseismic (e.g., mud slump on the continental slope, marine volcanic eruption) events also. Once a tsunamigenic event occurs, tsunami waveform data observed at offshore tsunami meters are represented by a linear superposition of the pre-calculated Green functions, and the estimated coefficients in turn represent a spatial distribution of initial tsunami source. The resultant tsunami waveforms at coastal points (i.e., tsunami waveform predictions) can be obtained by a linear combination of the estimated coefficients and the pre-computed Green functions. There are two kinds of offshore tsunami meter, one is a pressure sensor, and the other is GPS tsunami meter (Kawai et al. 2013). GPS tsunami meter measures vertical sea surface displacement relative to the reference station on land by applying real-time geodetic measurement technology, and the interpretation of the data is straightforward. On the other hand, a pressure sensor records not only a pressure change due to sea surface deformation by tsunami passage but also effects of coseismic permanent ocean bottom deformation, ocean bottom shaking by seismic wave passage, and pressure wave in the ocean water. In the Green function calculation, the correction to the coseismic ocean bottom deformation effect is taken into account (Tsushima et al. 2012), because for near-field tsunami forecast purpose, cases where tsunami meters are installed inside of the tsunami source area must be considered. And also, removal of pressure change due to ocean bottom shaking and pressure wave in the ocean water is essential (Matsumoto et al. 2012). A good performance of the tFISH is shown by its retrospective application to the tsunami events near Japan such as the 2011 Great East Japan Earthquake (Mw9.0) (Tsushima et al. 2011). The JMA is planning to incorporate this technique in the tsunami warning operation in the near future.

Tsunami Early Warning System in Japan

Before describing the present status of the JMA's TEWS, its historical review is briefly given.

The JMA started tsunami warning service in 1952. Since then, the JMA has been making effort to integrate its TEWS, and this is exactly a history of "fight against time."

At the commencement of the tsunami warning service, the JMA had 46 seismometers at meteorological observatories (basically near populated area on a sedimental layer). In case of earthquake occurrence, seismograms were read at observatories, and their results (P and S arrival times and maximum amplitudes) were transmitted to the headquarters in a telegram format. At the headquarters, hypocenter and magnitude were determined manually, and tsunami warning grade was determined by using empirical chart based on the relation between tsunami amplitude, earthquake magnitude, and epicentral distance to a coast of the past events. It took about 15–20 min to disseminate tsunami warning.

In the late 1960s to early 1980s, telemetry technology to transmit seismic waveform data from observatories to the headquarters and processing computer were introduced. Detection capability of the earthquake was improved by monitoring collected seismic waveforms at one place, and the P- and S-picking precision was also improved by introducing a digitizer. By these, dissemination time was reduced to 12–13 min.

In 1983, the Mid-Japan Sea Earthquake (Mjma7.7) occurred. The JMA disseminated tsunami warning in 14 min, but in about 7 min, tsunami struck the nearest coast, killing 104 people. After this event, the JMA deployed more sophisticated computer system for the seismic waveform processing. A graphical man-machine interface was introduced for more accurate and quicker phase reading and hypocenter and magnitude calculation. Still, the empirical method was used for the tsunami amplitude estimation. Dissemination time was reduced to 7–8 min.

In 1993, the Southwest-off Hokkaido Earthquake (Mjma7.8) occurred. The JMA disseminated a tsunami warning in 5 min, but the tsunami struck Okushiri Island in a few minutes,

and 230 people were killed or lost by tsunami. After this event, the JMA totally replaced the seismic network. All seismometers were installed in remote unmanned sites on hard rock, and the total number of the site increased to about 180. Dissemination time was reduced to 3–5 min.

In 1999, the JMA introduced a numerical tsunami propagation simulation technique, replacing the previous empirical regression formula, to improve tsunami forecast accuracy, and successfully disseminated tsunami warnings for 2003 Tokachi-Oki Earthquake (Mjma8.0) in 3 min after the earthquake occurrence. The observed tsunami amplitudes were in good agreement with forecast ones.

In 2011, the Great East Japan Earthquake (Mw9.0) occurred. The JMA disseminated the highest-grade “Major Tsunami Warning” in 3 min, but still, about 20,000 lives were lost. The JMA carefully examined the cause and summarized the problems as follows (Hoshiba and Ozaki 2012; Japan Meteorological 2013):

1. Magnitude underestimation: The JMA disseminated the first tsunami warning based on the quickly calculated Mjma7.9 in 3 min, and it was a significant underestimation. The JMA could not recognize a possibility of Mjma underestimation at the time of the warning dissemination.
2. The warning statement: The stated “estimated tsunami amplitude 3 m” may have delayed evacuation in coastal areas guarded with high breakwater.
3. Prompt Mw calculation failure and insufficient warning revision procedure based on offshore tsunami meter and the importance of the first warning: The JMA failed to revise the warning based on CMT and Mw solution in 15 min due to broadband seismic data overscale at most of the domestic stations. The JMA revised the tsunami warning in 28 min based on the observed tsunami amplitude of offshore GPS tsunami meter by applying empirically determined amplification coefficient (cf. Green’s law) to convert to a value at a coast, but could not utilize more offshore cable-type tsunami meter (pressure sensor) data, which detected

tsunami about 10 min earlier, for revision because the quantitative utilization method was not established yet. Moreover, the revised warning could not reach to the public in some areas due to power outage or communication linkage failure by tremendous strong shaking, and the first warning may have been the only message that the residents could get.

4. The tsunami information statement: The JMA disseminated a tsunami observation information in which “the initial wave amplitude 0.2 m” was mentioned. This was a fact, but it may have led to delays or interruptions in evacuation by giving residents an underestimating threat.

After the investigations on the JMA’s tsunami warning/information improvements in cooperation with intelligent persons, municipalities, broadcasting companies, and other relevant disaster mitigation organizations, JMA integrated observation network, dissemination procedure, and warning/information statements as described below.

The present status of the JMA’s TEWS is as follows.

Seismic Network

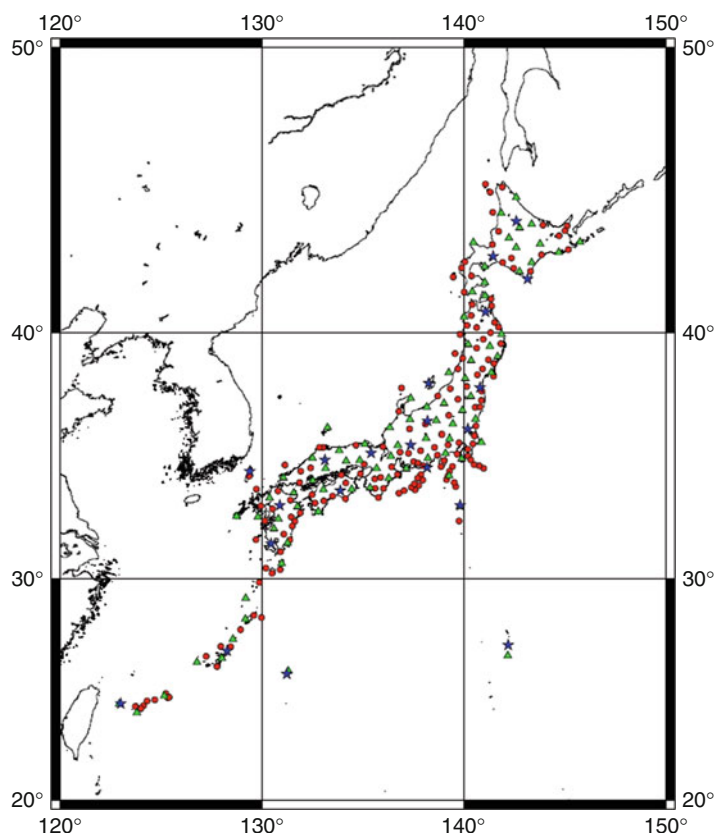
The JMA operates about 280 seismometers installed in Japan (Fig. 2). The seismic waveform data are sent to JMA Headquarters (Tokyo) and Osaka Regional Headquarters (as a backup system of the Tokyo Headquarters, to enable ceaseless tsunami warning dissemination even in case of the Tokyo Headquarters collapse) simultaneously and continuously on a real-time basis through dedicated telephone lines.

There are two kinds of sensors deployed at each station: a short-period velocity sensor and an accelerometer. The records of the short-period sensor are mainly used for precise picking of the onset times of P and S phases which are required for an accurate hypocenter determination. Accelerometer records are used for the calculation of magnitude of large earthquakes in the case that the ground motion amplitude exceeds the dynamic range of the short-period sensor.

STS-2 broadband seismometers have additionally been installed at 20 stations. These broadband velocity seismic records are mainly

Tsunami Forecasting and Warning,

Fig. 2 Seismic network of the JMA used for tsunami forecasting. *Red circles*, *green triangles*, and *blue stars* denote the locations of seismic stations operated by JMA. The average spacing between the about 280 stations is 40–50 km. *Blue stars* and *green triangles* denote the location of stations where STS-2 velocity broadband seismometers and broadband strong motion meters have been installed, respectively, in addition to the Japanese short-period velocity-type seismometers and accelerometers



used for obtaining CMT solutions and moment magnitudes (M_w). But, broadband seismometer records are prone to saturate for strong shaking like in the case of the Great East Japan Earthquake in 2011, so the broadband strong motion meters were additionally installed at 80 stations after this event to assure the tsunami warning updates based on CMT and M_w solutions in about 15 min even for nearby gigantic earthquakes.

At all stations, battery backup ability (up to 72 h) and satellite communication link to backup ground lines were reinforced also.

Real-Time Seismic Data Processing System

The JMA applies the short-term average/long-term average (STA/LTA) method as trigger criteria. It is most widely adopted in the seismological community and used to limit seismic recordings to relevant seismic event waveforms.

When the ratio of STA/LTA exceeds the trigger threshold, it is supposed that a significant signal, possibly an earthquake signal, has been detected by the considered station.

STA/LTA trigger may occur due to some troubles or local noise at one station. To avoid “false trigger,” the JMA also applies other criteria including the group trigger criterion. A group is set to include several neighboring stations. When the STA/LTA trigger turns on at a certain number of stations in one group, the system considers it to be caused by an earthquake.

For the precise picking of the P onset, an autoregressive (AR) model is applied to represent seismogram time series. Akaike’s information criterion (AIC) is used to search the optimal time to separate the seismogram into two sub-stationary time series (noise part and signal part) which are represented by AR models, respectively, in a certain time window around trigger time (Takanami and Kitagawa 2002; Yokota et al. 1981).

The usual least square method is used in the hypocenter calculation. As for the velocity structure, the Japan local standard model JMA 2001 (Ueno et al. 2002) is applied. To better locate the hypocenter, S arrival times are used in addition to P arrival times. First, a preliminary hypocenter is calculated by using only P arrival times picked by an automatic picker. Then, S arrival times are picked automatically by applying the AR model and AIC in a time window centered at the theoretical S arrival times estimated from the preliminary hypocenter location and origin time. The final hypocenter is then calculated by using both the P and S arrival times.

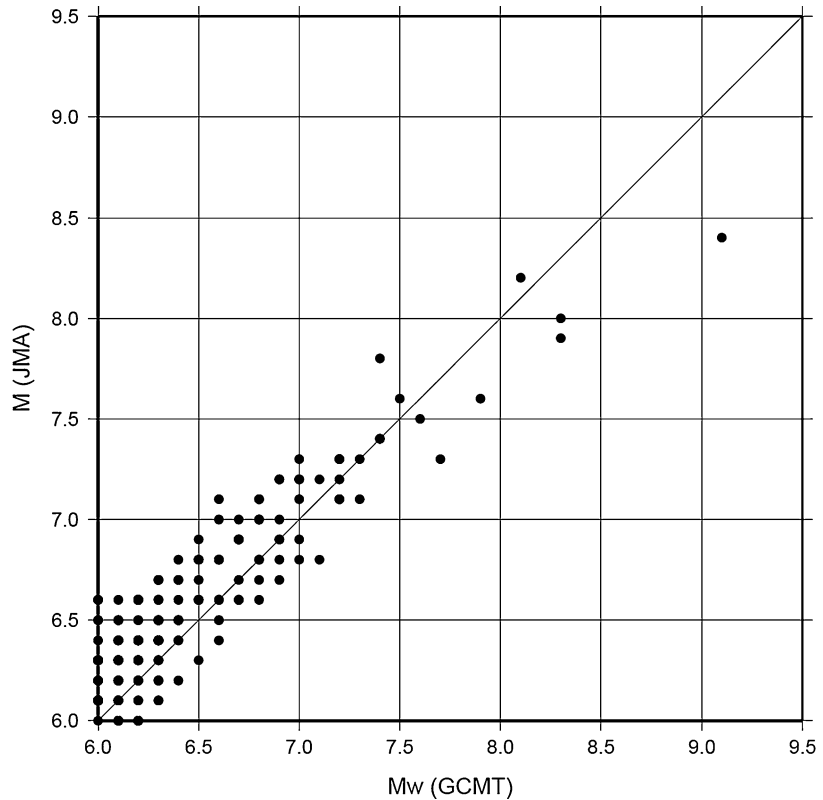
For magnitude calculation, one of the local magnitude definitions in Japan, the JMA magnitude (M_{jma}) according to Katsumata (2004) is used, in which average amplitude decay characteristic in Japan is taken as a decay correction term of the formula. Maximum amplitude of the horizontal ground displacement is the measured input value in the formula. The displacement

waveform is calculated by applying real-time recursive filter that integrates twice the accelerometer data and applies a high-pass filter to avoid computational instability. Due to historical reasons, the long-period cutoff of the high-pass filter is set at 6 s. Since unsaturated strong motion data is used, magnitude calculation can be started promptly after the earthquake occurrence by using data from stations very close to the epicenter. Therefore, such magnitude data is suitable for the tsunami warning purposes of local events.

The comparison between M_{jma} and the moment magnitude M_w of the Global CMT Project's solution is shown in Fig. 3. M_{jma} values are on average comparable with M_w for earthquakes with magnitudes between $M_{6.0}$ and about 8.0, but saturates for events with magnitudes over 8.0. M_{jma} is superior in the promptness and M_w is superior in representing tsunamigenic potential even for gigantic or tsunami earthquake, and they are complementary to each other. So, the JMA's strategy for tsunami warning is to disseminate the

Tsunami Forecasting and Warning,

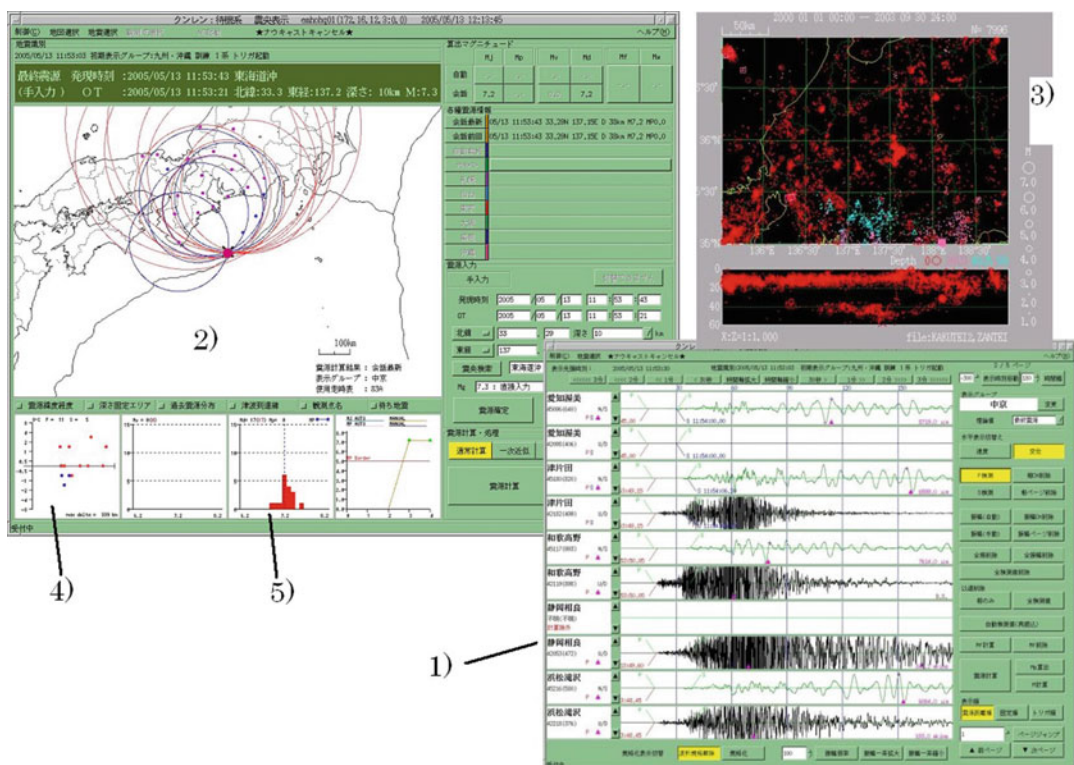
Fig. 3 Relation between M_w (Global CMT) and M_{jma} , based on events that occurred after September 2003 in and around Japan. Both magnitudes agree on average well up to around magnitude 8.0, but M_{jma} tend to saturate for magnitude over 8.0



first warning in 3 min based on Mjma and revise it, if necessary, based on Mw in 15 min (as for the revision based on Mw, see subsection “Recent Improvements”).

All procedures mentioned above are run automatically by the computer. Considering the possible impact of a tsunami warning on the potentially affected communities, a human check of the adequacy of the calculated results is indispensable for minimizing false alarms. The following items are checked by the operators, using a man-machine interface.

1. Adequacy of the phase pickings and maximum amplitude readings: The seismic waveforms are plotted together with picked and theoretical arrival times for P and S and the marked position at which the maximum amplitude was measured. This plot also allows them to discriminate noise from seismic signals.
2. Adequacy (mathematical) of the hypocenter location: A hypocenter plot map is used in which the locations of used seismic stations and the circles denoting the calculated epicentral distance from each station are shown. If the circles intersect densely at one point, and if the used stations assure a wide azimuthal coverage around the source, then the calculation is judged as fine.
3. Adequacy (seismological) of the hypocenter location: The hypocenter plot on the background seismicity map is used together with the vertical cross section. If the hypocenter is located in a region where no background seismicity is found, it should be reexamined carefully.
4. Adequacy of the depth estimation: Travel time residuals are plotted as a function of epicentral distance. If the residuals depend on epicentral distance, the depth estimation should be revised.



Tsunami Forecasting and Warning, Fig. 4 Examples of JMA's man-machine interface screen image in the seismic data processing system. Numbers in the figure are related to the respective numbers of items explained in the main text

5. Adequacy of the magnitude calculation: A histogram of station magnitudes is used. Outliers are checked and excluded in the recalculation.

Sample images of the man-machine interface are shown in Fig. 4

The JMA's man-machine interface is designed so that re-picking and recalculation of hypocenter and magnitude can be done within about 30 s.

In addition to the adequacy check described above, the possibility of *Mjma* underestimation is also checked, incorporating what the JMA has learnt from the 2011 Great East Japan Earthquake (see subsection "[Recent Improvements](#)"). In case of probable underestimation of *Mjma*, the preset maximum magnitude of the area, inferred from up-to-date seismological studies, is adopted instead of calculated *Mjma* to avoid an

underestimation of tsunami amplitude in the first issuance of warning, to save lives even in the case in which revised warnings do not reach to the public due to power failure. Figure 5 shows the flowchart of the tsunami warning dissemination procedure, including its revision.

Calculated hypocenter parameters (latitude, longitude, depth, and magnitude) are transferred to the following tsunami forecast system as its input.

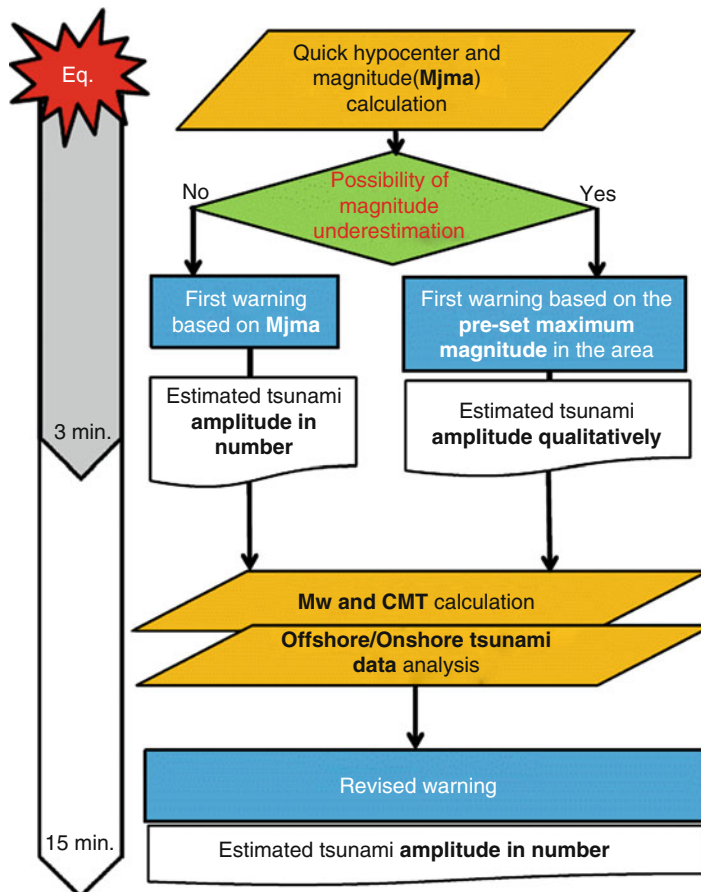
Tsunami Forecast System

Tsunami Forecast in Japan

For a tsunami forecast to be useful, a mere general description of the tsunamigenic potential estimated from calculated hypocentral parameters is not sufficient. From the issued message, the recipient disaster management organization should be able

Tsunami Forecasting and Warning,

Fig. 5 Flow chart of the JMA's tsunami warning dissemination procedure including its revision. The possibility check of *Mjma* underestimation was newly added in the flow after the 2011 Great East Japan Earthquake



to clearly grasp the severity of the expected disaster in its own jurisdiction. In Japan, a tsunami forecast is disseminated to each coastal block (the coast line of Japan is separated into certain number of blocks), giving three kinds of grade corresponding to the anticipated severity of the disaster.

Tsunami Forecast to Coastal Block The coastal blocks divide the Japanese coasts into 66 regions (Fig. 6). These blocks are defined by taking into account:

- (a) The administrative districts of local governments, which are the disaster management units to take actions in emergency situations
- (b) The sea areas to which each coast is facing
- (c) The uniformity in the behavior of tsunami
- (d) The precision of tsunami forecast technique

Basically, one coastal block corresponds to one prefecture or finer.

Before the introduction of numerical tsunami simulation technique, the total number of coastal blocks was 18, due to a low precision of empirical tsunami amplitude estimation method.

Tsunami Forecast Criterion and Category The JMA categorizes tsunami forecast into “Major Tsunami Warning,” “Tsunami Warning,” and “Tsunami Advisory.”

In Japan, there exist ample number of materials on the relation between tsunami disaster and observed tsunami amplitude along the coast including those for the 2011 Great East Japan Earthquake (City Bureau, Ministry of Land, Infrastructure, Transport and Tourism 2011). From the materials, high correlation between these two can be seen. Therefore, estimated maximum tsunami amplitude at the coast is used as a criterion of the tsunami forecast in Japan. The speed of water current in the sea near the coast could be another good index for assessing the tsunami disaster potential. But, up to now, only a few examples of the current observation are available. Therefore, the estimated tsunami amplitude at the coast is still the most appropriate criterion for tsunami forecast.

In Japan, tsunami disaster in the land area occurs when the tsunami amplitude exceeds 1 m (Shuto 1998), so the “Tsunami Warning” criterion is set at above 1 m. When tsunami amplitude exceeds 3 m, the proportion of damaged wooden houses and fishing boats increases remarkably (Hatori 1984; Shuto 1991, 1992). Therefore, the JMA issues the highest warning grade “Major Tsunami Warning” when the estimated maximum tsunami amplitude is above 3 m. If a “Tsunami Warning” or “Major Tsunami Warning” is issued, the head of municipality must order evacuation to the residents in the “Tsunami Evacuation Zone” assigned by each municipality in advance. It is recommended to set the possible maximum evacuation (inundation) zone corresponding to “Major Tsunami Warning” and the evacuation zone where inundation is expected for 3 m amplitude tsunami strike corresponding to “Tsunami Warning.”

Even when the tsunami amplitude is less than 1 m, it is occasionally the case that bathing people are affected and aquaculture rafts for marine products industry are damaged. Considering safety criteria in sea-bathing places (beaches, bays), and the relation between the tsunami amplitude near the coast and damage on aquaculture rafts in the past, the JMA’s criterion for “Tsunami Advisory” is set at 20 cm. In the case that a “Tsunami Advisory” is issued, there is no need of evacuation in the land areas, except vulnerable very low-land areas, but people on beaches or swimming should go to higher places.

Table 1 shows the tsunami forecast grades and classifications of the estimated maximum tsunami amplitude that are used by the JMA when issuing tsunami information for each coastal block after a tsunami forecast has been made. The JMA amended the expressions and number of the classes of estimated tsunami amplitude in March 2013 reflecting what JMA has learnt from the 2011 Great East Japan Earthquake as follows:

1. Usage of qualitative expression

In case that preset maximum magnitude is adopted (i.e., possible Mjma underestimation case), considering a significant uncertainty of

Tsunami Forecasting and Warning, Table 1 Tsunami forecast grades and expressions of corresponding classifications of estimated maximum tsunami amplitudes as used in tsunami information issued by the JMA for each

Category of Warning/Advisory	Before Estimated Tsunami Amplitude	Current(2013.Mar -)		
		Expression		Tsunami Amplitude Classification
		Quantitative	Qualitative	
Major Tsunami Warning	Over 10 m, 8 m, 6 m, 4 m, 3 m	Over 10 m	Huge	10 m<Amp.
		10 m		5 m<Amp.<=10 m
		5 m		3 m<Amp.<=5 m
Tsunami Warning	2 m, 1 m	3 m	High	1 m<Amp.<=3 m
Tsunami Advisory	0.5 m	1 m	(None)	20 cm≤Amp.≤1 m

coastal block after a tsunami forecast has been made. Current (after 2013 Mar.) and previous expressions are shown

the magnitude and to transmit the maximum emergency to the public, estimated tsunami amplitude is stated only qualitatively, namely, “Huge” for Major Tsunami Warning and “High” for Tsunami Warning, based on the consideration that warnings must urge the residents to evacuate.

2. Reduction of the number of classes

In case of no possibility of Mjma underestimation or after Mw and CMT solutions are available, quantitative expression (number) is used for estimated tsunami amplitude. The number of classes is simplified from 8 (0.5 m, 1 m, 2 m, 3 m, 4 m, 6 m, 8 m, 10 m, or over) to 5 (0.2–1 m, 1–3 m, 3–5 m, 5–10 m, over 10 m) considering that scatter of tsunami amplitudes in one forecast block becomes large as estimated/observed amplitude becomes large, as well as a feasibility of multiple evacuation zones setting. And in the information statement, upper bound for each class (e.g., “3 m” for 1–3 m class) is mentioned from a psychological point of view (a number is regarded as “reference value” to judge the proper action for individuals).

When the estimated tsunami amplitude is less than 20 cm, no warnings or advisories are issued, but a message “No threat of tsunami disaster” is promptly provided to the public in order to cancel

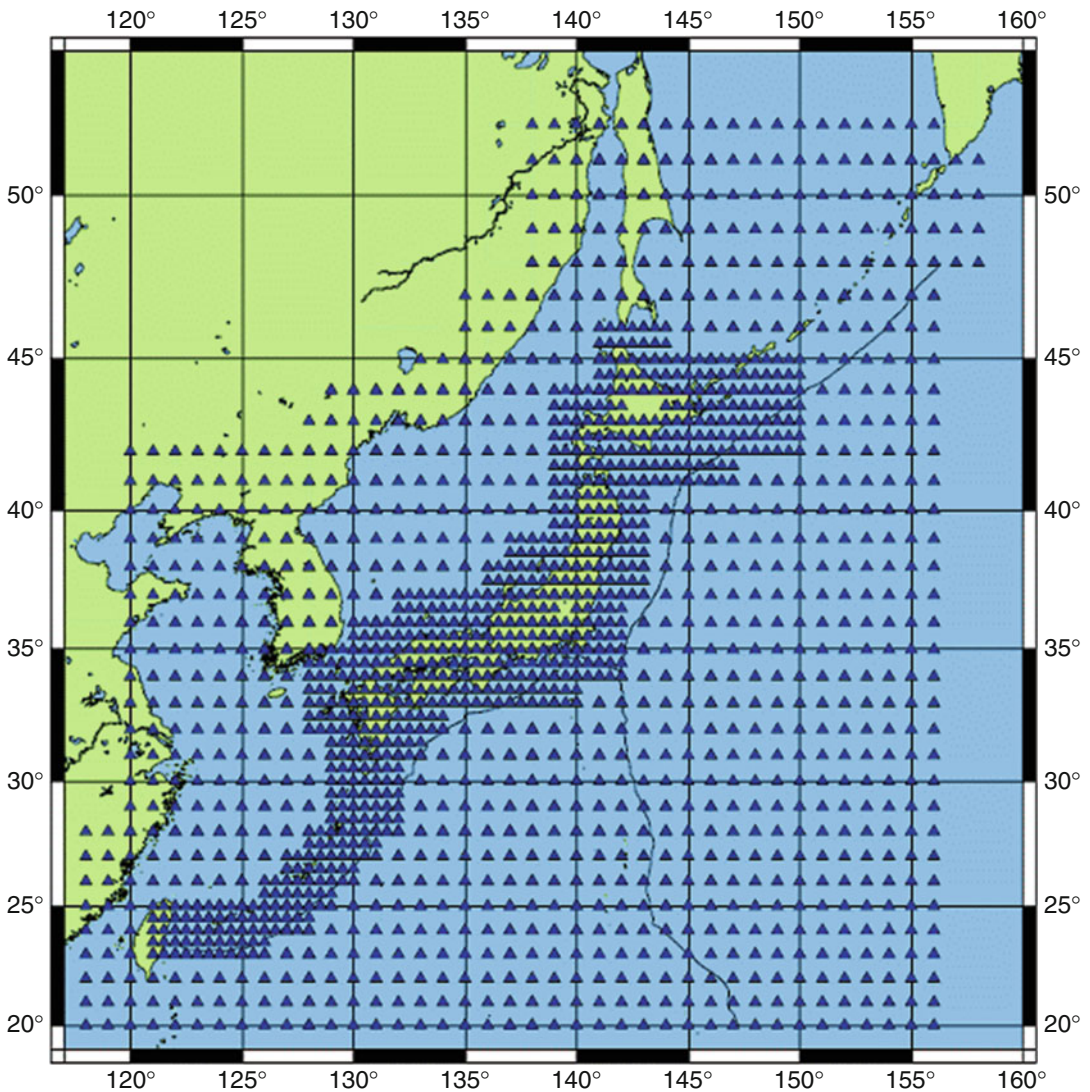
unnecessary spontaneous evacuations in accordance with the widely spread recommendation in Japan: “When you feel a strong shaking near a coast, run to a high place without waiting for tsunami forecasts from the JMA.”

Tsunami Database Creation The JMA introduced in 1999 a numerical simulation technique to implement accurate tsunami warning. However, even most advanced computers require substantial time for the completion of the calculations. Therefore, it is impossible to issue prompt tsunami warning based on numerical simulation technique even if the calculations start simultaneously with the occurrence of an earthquake.

Alternatively, the JMA employs a database method to achieve a breakthrough on this issue. Tsunami propagation originating from various locations, fault sizes, and likely rupture mechanisms is calculated in advance, and the results, namely, estimates of maximum tsunami amplitudes and arrival times, are stored in a database together with the associated hypocenter location and magnitude.

If an earthquake occurs, the most appropriate case for the actual hypocenter location and magnitude is retrieved from the database, and the tsunami forecast is issued accordingly.

Hypocentral Location The first step to create a tsunami database is to set fault models. “Simulation



Tsunami Forecasting and Warning, Fig. 7 Distribution of simulation points in and around Japan. *Solid triangles* denote locations of simulation points to which

likely hypothetical fault models are attributed. Spacing between the simulation points is 0.5° in near coastal areas and 1.0° in offshore areas

points” are defined as surface projection locations of the center of possible causing faults and are placed in the areas where tsunamigenic earthquakes are likely to occur. These points are selected on the basis of background seismicity and the records of the past earthquakes that generated tsunami.

Figure 7 shows the locations of simulation points in Japan and surrounding areas.

The total number of scenarios calculated for these points is about 64,000 (lat. $\times$ lon. $\times$ dep. $\times$

mag.). Earthquakes with hypocenter depth of more than 100 km or magnitude less than 6.2 are not considered, because they do not cause disastrous tsunami (Okada and Tanioka 1998).

Fault Parameter Setting Figure 1 shows fault parameters used in the numerical simulation. Length L , width W , and slip amount D of the fault are represented in terms of empirical formulas as functions of magnitude. The following are

common relations used in Japan (Utsu et al. 2001) based on $M = M_{jma}$:

$$\log L = 0.5 M - 1.8 \text{ with } L \text{ in km}$$

$$W = L/2 \text{ or } \log W = 0.5 M - 2.1 \text{ with } W \text{ in km}$$

$$\log D = 0.5 M - 3.3 \text{ with } D \text{ in m}$$

These formulas are consistent with the next two formulas, assuming a common rigidity $\mu = 2.0 \times 10^{10} \text{ N/m}^2$ of the rock material in the source area:

$$M_0 = \mu LWD$$

$$\log M_0 = 1.5 M_w + 9.1 \text{ (in international standard units, i.e., } M_0 \text{ in Nm = Newton meter)}$$

The dip (δ), strike (ϕ), and slip (λ) angles of the fault at each simulation point are based on average values of past earthquakes at or near these locations. But if they are uncertain, they are assumed to be those of a pure reverse fault ($\lambda = 90^\circ$) whose strike is parallel to the trench or nearby coast and with a dip angle of 45° , corresponding to an efficient tsunami generation in view of disaster management.

Initial Value for Numerical Simulation of Tsunami Propagation The vertical deformation of sea floor due to fault motion is calculated by the elastic theory (Okada 1985), using fault parameters as set above. The initial deformation of the sea surface is assumed to be identical to the vertical deformation of the sea floor and used as input value for the numerical simulation, because in most cases, rupture propagation velocity is much higher than the tsunami propagation velocity, and the ocean bottom deformation can be treated to occur instantaneously (Kajiura 1970).

Numerical Simulation of Tsunami Propagation Nonlinear long-wave approximation equation with advection term and ocean bottom friction term is adopted as the equation of motion and solved together with the equation of continuity as shown below by finite difference method with staggered leap-frog scheme (Satake 1995).

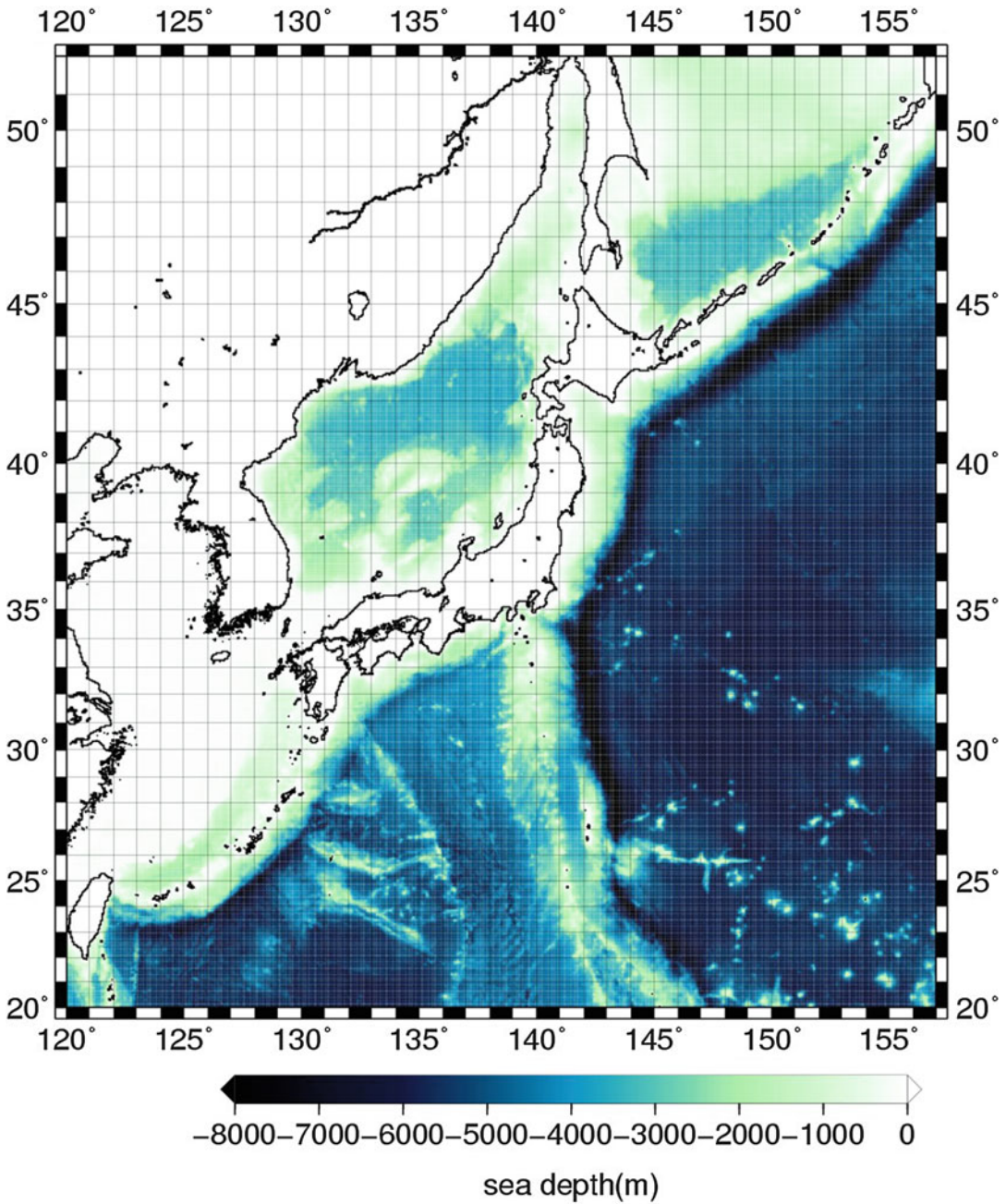
$$\begin{aligned} & \frac{\partial V_x}{\partial t} + V_x \frac{\partial V_x}{\partial x} + V_y \frac{\partial V_x}{\partial y} \\ &= -fV_y - g \frac{\partial h}{\partial x} - C_f \frac{V_x \sqrt{V_x^2 + V_y^2}}{d+h} \\ & \frac{\partial V_y}{\partial t} + V_x \frac{\partial V_y}{\partial x} + V_y \frac{\partial V_y}{\partial y} \\ &= fV_x - g \frac{\partial h}{\partial y} - C_f \frac{V_y \sqrt{V_x^2 + V_y^2}}{d+h} \\ & \frac{\partial h}{\partial t} + \frac{\partial}{\partial x} \{V_x(h+d)\} + \frac{\partial}{\partial y} \{V_y(h+d)\} = 0 \end{aligned}$$

where V_x and V_y are the x (east) and y (south) components of water particle motion velocity averaged in the vertical direction, and h and d are sea surface displacement and sea depth, respectively. f is the Coriolis parameter ($2\Omega \cos \theta$, where Ω is the angular frequency of the Earth's self rotation and θ is the colatitude), and C_f is the sea bottom friction coefficient.

Advection and ocean bottom friction are considered only in sea areas with sea depth less than 100 m. Coriolis force is considered only in the distant tsunami case (see subsection “[Tsunami Forecast for Distant Event and Northwest Pacific Tsunami Advisory \(NWPTA\)](#)”).

Figure 8 shows the computational area. Mesh size is 1 arc minute throughout the computational area. Eight hours of propagation simulation is conducted for all scenarios with a fixed time interval of 3 s for the integration. The Coriolis force is not considered for local and regional tsunami simulations. Total reflection boundary condition is used at the land-ocean boundary, and open boundary condition is used outside of the computational area.

To represent the tsunami waveform correctly in a shallow sea area, very fine-mesh bathymetry data is necessary (in a strict sense, 20 or more grid points are necessary within one wavelength (Shuto et al. 1986)), and a vast time is required for the completion of such detailed calculations. To overcome this difficulty, the numerical simulation with the long-wave approximation is applied only to points which are a few to a few 10 km seaward from the coast (“forecast points”) where sea depth is about 50 m. Then, tsunami amplitude at the coast is calculated by using Green's law described in the next section.

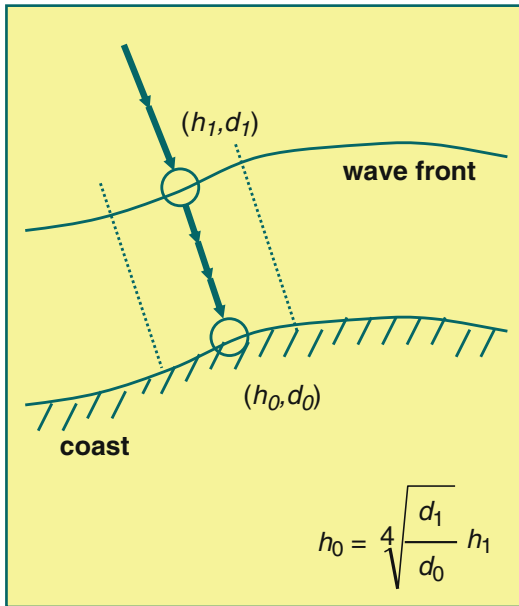


Tsunami Forecasting and Warning, Fig. 8 Computational area for creating the Japanese tsunami simulation database. Bathymetry is given by a color grade scale. One

arc minute mesh bathymetry data is used throughout this area for the tsunami simulation

Derivation of Tsunami Amplitude at Coast To estimate tsunami amplitude at the coast, Green's law (energy conservation law) is applied to the estimated amplitude at the forecast points just

seaward of the coast. As shown in Fig. 9, Green's law says that the tsunami amplitude at the coast is represented by the fourth root of the ratio of sea depth at the forecast point and at the coast:



Tsunami Forecasting and Warning, Fig. 9 Schematic sketch to illustrate the application of Green's law for tsunami amplitude calculation in near coastal areas. Ray convergence and divergence are ignored

$$h_0 = \sqrt[4]{\frac{d_1}{d_0}} h_1$$

where h and d denote tsunami amplitude and sea depth, and suffixes "0" and "1" denote the value at the coast and forecast point.

As the forecast points are set close enough to the coast, horizontal convergence and divergence of the rays can be ignored, i.e., the wave front is regarded as almost parallel to the coast due to a refraction toward the direction of the largest sea-depth gradient. As for the " d_0 " value, the sea depth at the coast, 1 m is assumed, but actually measured value would be more appropriate, if available.

Adequacy of Green's Law Application The adequacy of Green's law application to estimate tsunami amplitude at the coast is examined by using the TSUNAMI-N2 (Imamura 1997) program. It is equivalent to the program used at the JMA (Satake 1995) to create tsunami simulation database. Up to 150 m mesh at the finest was used by adopting nesting technique, and the tsunami

amplitudes estimated at the coastal mesh are considered "true" ones. We compared the case using 1,350 m mesh throughout the computational area (which is equivalent to 1 arc minute mesh) and then estimating the tsunami amplitude at the coast by applying Green's law to the value at the offshore forecast point, with the expected "true" amplitude at the coastal point using finer mesh. Forecast points actually used for the database creation are shown in Fig. 10. Hypothetical fault models corresponding to two different magnitudes (8.0 and 6.8) have been depicted in this figure as red and violet rectangles, respectively, south off the Pacific coast of Hokkaido Island. The results are shown in Fig. 11a, b. For both events the estimated tsunami amplitudes at the coast by using a 1,350 m mesh and applying Green's law to the forecast points (solid squares) agree well with the means (open squares) and medians (open circles) of tsunami amplitudes calculated by using a finer mesh in each coastal subsection. The coastal subsections are separated by middle points between the forecast points (see upper part of Fig. 10). Thus, it was confirmed that the application of Green's law to tsunami amplitudes at the offshore forecast points yields amplitude estimates at the coast that are comparable with those calculated with much more time-consuming fine-mesh simulations. Therefore, we consider the former as representative estimates of the tsunami amplitude in the coastal subsection centered at the coastal projection of the forecast point.

It could also be shown that direct estimates of tsunami amplitudes at the coast based on coarse-mesh simulations will underestimate values. This trend is even clearer for smaller event with shorter typical tsunami wavelengths (Fig. 11a). Therefore, it is not proper to estimate the tsunami amplitude directly at the coastal mesh when the mesh size is coarse relative to the typical tsunami wavelength.

However, there will be some issues to be considered in the future in this method as follows:

- (a) Strictly speaking, Green's law should only be applied to a direct wave. Therefore, in case the maximum amplitude is created by the

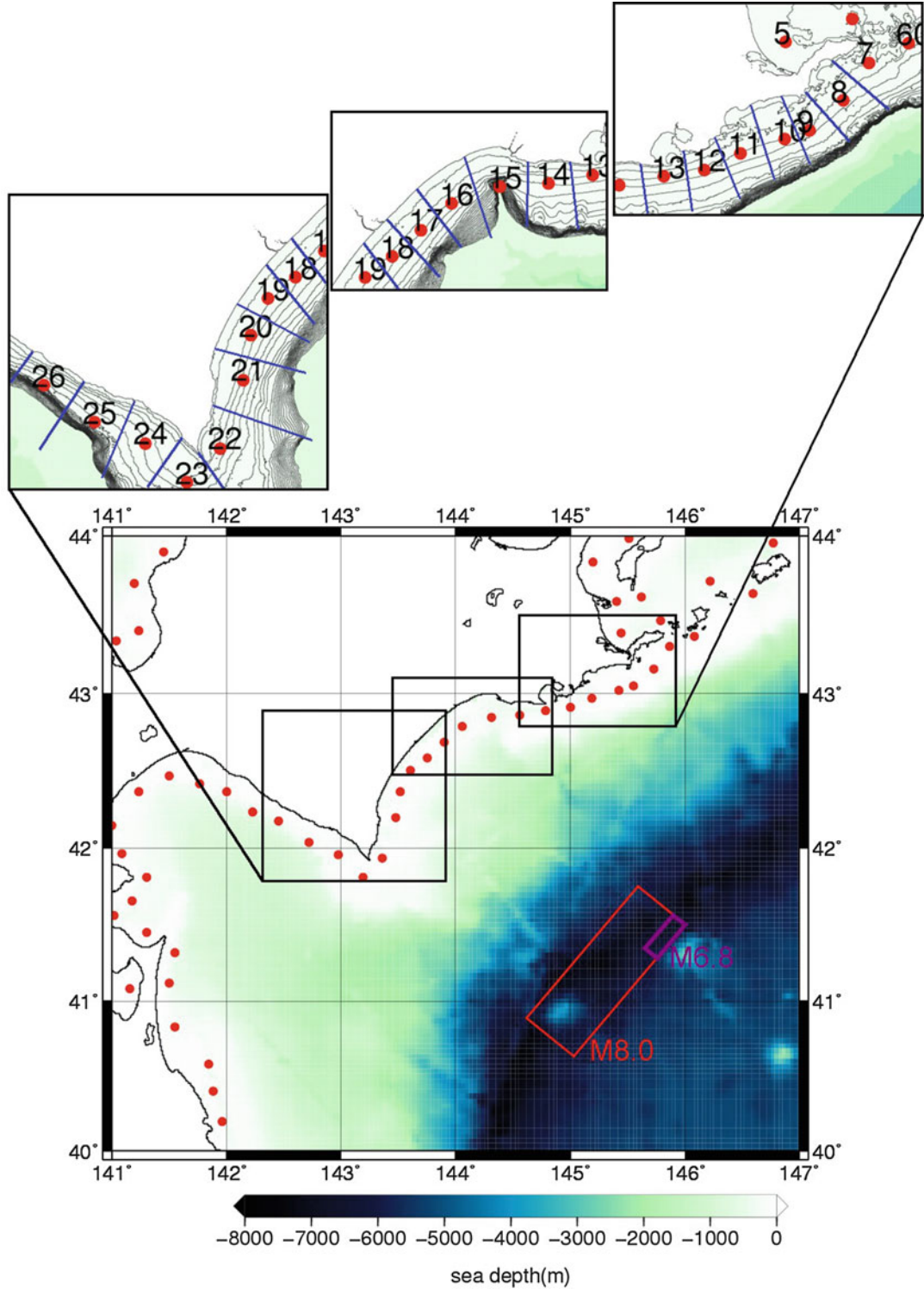


Fig. 10 (continued)

superposition of multiply reflected waves, then this method can give overestimated tsunami amplitude.

- (b) Scatter of estimated tsunami amplitude in one coastal subsection differs from section to section depending on the complexity of the coast line feature. It would be very difficult to precisely represent tsunami amplitudes for individual coastal points even if a very fine bathymetry data is used in a simulation, because a short-wavelength tsunami is significantly affected by small-scale bathymetry and coastal feature or by a small change in an initial tsunami waveform. Therefore, incorporating a standard deviation of estimated tsunami amplitude to represent a degree of scatter in one coastal subsection, as well as a median or mean, in the tsunami warning/advisory grade determination would be effective for disaster management.

Retrieval of the Most Appropriate Case from the Database

In general, no identical case for actually determined latitude, longitude, hypocenter depth, and magnitude exists in the database. Thus, some kind of selection rule is necessary to choose the most appropriate one.

Hypocenter Depth and Magnitude The logarithm of the estimated tsunami amplitude changes approximately positively linear with magnitude and negatively linear with hypocenter depth. Therefore, first, one has to calculate the logarithm of the estimated tsunami amplitudes for four cases (i.e., for two different depths and magnitudes each) which include the calculated depth and magnitude in between, then apply a two-dimensional linear interpolation, and finally

convert the log values back to a tsunami amplitudes.

Latitude and Longitude As for the location of the epicenter, either one of the two methods is used depending on the hypocentral area.

Interpolation Method

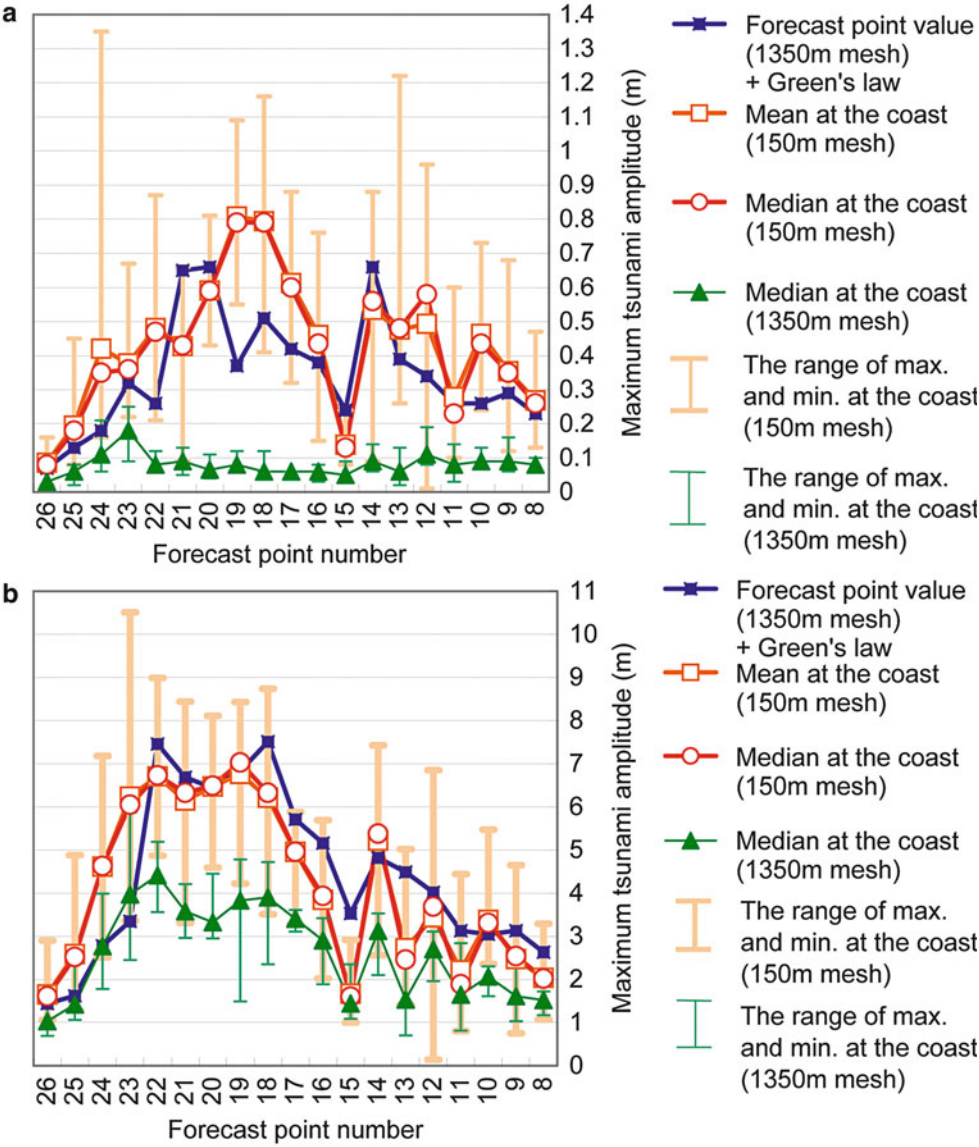
One selects the nearest four locations of simulation points in the database which enclose the epicenter determined for the real event and performs a two-dimensional linear interpolation. This method is used in far off the coast areas where horizontal shifts of the source location cause relatively smooth changes in the estimated tsunami amplitude at the coast. This is applicable in areas with a simulation point spacing of 1° (see Fig. 7).

Maximum Risk Method

In the case that the assumed fault locations in the database differ from the real location, tsunami amplitude estimates near the epicenter may differ substantially from observed ones for events occurring near to the coast. The uncertainty in the relative location of the epicenter on the surface projection of the seismic fault has to be considered at an early stage of the seismic data analysis, and all possible cases should be taken into account. Four cases with the epicenter located on one of the four corners of the fault are the most extreme cases, and the truth lies in between. Therefore, in the case of near coastal sources and with a view to disaster management, cases corresponding to every simulation point inside the rectangular area shown in Fig. 12 (solid circles) have to be considered. This rectangular area is centered at the epicenter (star mark). Its length and width are determined via the

Tsunami Forecasting and Warning, Fig. 10 The map shows the distribution of forecast points (*red dots*) along the southern and eastern shore of Hokkaido and part of NE Honshu together with the surface projections of two hypothetical faults representing earthquakes with magnitude 8.0 (*red rectangle*) and 6.8 (*violet rectangle*), respectively. The depth of the fault top is 1 km from the sea bottom for both cases. Pure reverse fault with dip angle 45° is

assumed. In the upper half of the figure, the coastal blocks are enlarged, showing the locations of forecast points denoted by numerals and the *blue separation lines* between subsections of the coast. The numbers correspond to the “forecast point number” given on the abscissa of Fig. 11. Also depicted are bathymetry *contour lines* near the coast at 20 m depth intervals

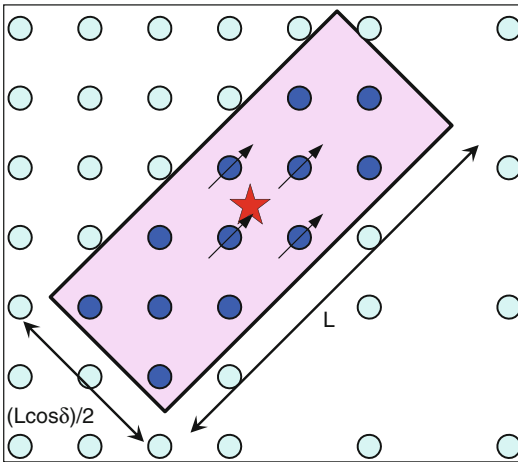


Tsunami Forecasting and Warning, Fig. 11 The presented diagram illustrates the adequacy of Green's law application. Shown are the tsunami forecast results for the hypothetical earthquakes of magnitude 6.8 (*above*) and 8.0 (*below*). On the abscissa, the numbers of the "forecast points" depicted in Fig. 10 are given. On the ordinate the maximum tsunami amplitudes in meter are given, as they result from the application of different

modeling procedures (fine mesh as well as course mesh with and without applying Green's law; see legend). *Vertical thick bars* denote the range between maximum and minimum values of estimated tsunami amplitudes in each subsection using an up-to-150 m mesh. *Vertical thin bars* denote the range of directly estimated maximum and minimum tsunami amplitudes at the coast in each subsection using a 1,350 m mesh

empirical formulas given in subsection "Fault Parameters Setting." Note, however, that the projected width in Fig. 12 is only $0.5 L \cos \delta$ and depends on the source dip. Also one has to

remember that each simulation point is located in the center of the respective scenario fault in the database. Therefore, the length and width of the rectangle to select all possible "simulation



Tsunami Forecasting and Warning,

Fig. 12 Maximum risk method: The *star* denotes the epicenter and the circles the locations of simulation points. The *rectangle* delimitates the searching area from which the simulation points have to be retrieved. This *rectangle* is centered at the epicenter, and its length and width are L and $0.5 L \cos \delta$, respectively, where L is given by the scaling law (see subsection “[Fault Parameters Setting](#)”). The strike of the length is the same as that given to the nearest simulation point to the epicenter. All simulation points inside of the rectangle (*solid circles*) are selected. *Small arrows* point into the strike direction assigned to simulation points

points” are L (not $2 L$) and $0.5 L \cos \delta$, respectively, and the strike of the length is that given to the simulation point (denoted by an arrow) which is closest to the determined epicenter. And the worst case is chosen among them for each of the forecast points.

When readings of the tsunami arrivals at tidal gauges or tsunami meter become available, improbable simulation points can be excluded by applying inverse refraction diagrams. Thus, the uncertainty in the spatial extent of the tsunami source area can be reduced.

If the number of the simulation points inside of the rectangular area is less than 4, the nearest 4 simulation points from the epicenter are retrieved.

Tsunami Arrival Time Estimation Forecast points are located not along the coast but offshore. Travel time estimation by adding the travel time from the source area to the forecast point to

that from the forecast point to the coastal point may lead to substantial errors, especially when the tsunami source area is close to the coast. Therefore, for the tsunami arrival time estimation at the coastal point, an inverse refraction diagram from the coastal point to the source area is used to reduce such errors. Finite spatial extent of the source area, depending on the calculated magnitude value, is considered, and the earliest arrival time corresponding to the crossing point of the inverse refraction diagram and the outer rim of the source area is used with a view to disaster management.

Tsunami Forecast Assembling

The tsunami forecast grade for a considered coastal block depends on the maximum of the expected tsunami amplitudes at the forecast points located in that block which are converted to amplitudes at the coast by applying Green’s law. Forecast points are placed offshore parallel to the coast with spacing of about 20 km. The average number of forecast points in one coastal block is 9. As mentioned above, expected tsunami amplitude at a forecast point, including Green’s law application when using 1 arc minute bathymetry mesh, agrees reasonably well with expected mean or median tsunami amplitudes on the coast that have been calculated by using very fine bathymetry mesh in a coastal subsection separated in the middle between forecast points. This means that the expected tsunami amplitude at the forecast point can be regarded as being representative for the respective coastal subsection. We think this approach is reasonable with a view to disaster management requirements because the forecast grades for a coastal block are based on the maximum of these representative values in each coastal subsection. This prevents that the forecast grade is affected by abnormally large local maxima as they may result from very fine-mesh simulations (see Fig. 11). However, JMA informs in its public information releases that tsunami amplitude might very locally be much higher than in the forecast.

The content of tsunami warning/advisory is very simple, namely, tsunami grades and corresponding coastal block names. This is to

Tsunami Forecasting and Warning, Table 2 Tsunami forecast/information examples of JMA when the current expressions are applied to the Kuril event case in Nov. 15, 2006. Original texts are in Japanese and these are English translations. (a) Tsunami forecast of the first issuance (b) Tsunami information (estimated tsunami arrival

time and amplitude) that follows (a) (c) Tsunami information (estimated high tide and tsunami arrival times) that follows (a)(d) Tsunami information (observed results) (e) Tsunami forecast revision (f) Tsunami forecast cancellation

(a) Tsunami Forecast of the first issuance

Tsunami Warning

Issued by the Japan Meteorological Agency (JMA)

Issued at 2029JST 15 Nov 2006

"Tsunami Warning"

Eastern Part of Pacific Coast of Hokkaido

Okhotsk Sea Coast of Hokkaido

"Tsunami Advisory"

Central Part of Pacific Coast of Hokkaido

Western Part of Pacific Coast of Hokkaido

Northern Part of Japan Sea Coast of Hokkaido

Pacific Coast of Aomori Pref.

Iwate Pref.

Miyagi Pref.

Fukushima Pref.

Ibaraki Pref.

Kujukuri and Sotobo Area of Chiba Pref.

Uchibo Area of Chiba Pref.

Izu Islands

Sagami Bay and Miura Peninsula

Shizuoka Pref.

(b) Estimated Tsunami Arrival Time and Amplitude

Tsunami Information (Estimated Tsunami Arrival Time and Amplitude)

Issued by the Japan Meteorological Agency (JMA)

Issued at 2030JST 15 Nov 2006

----- Estimated Tsunami Arrival Time and Amplitude -----

Coastal Block

Arrival Time Amplitude

"Tsunami Warning"

Eastern Part of Pacific Coast of Hokkaido

2110 15 Nov 3m

(continued)

Tsunami Forecasting and Warning, Table 2 (continued)

Okhotsk Sea Coast of Hokkaido	2120	15 Nov	3m
"Tsunami Advisory"			
Central Part of Pacific Coast of Hokkaido	2130	15 Nov	1m
Western Part of Pacific Coast of Hokkaido	2150	15 Nov	1m
Northern Part of Japan Sea Coast of Hokkaido	2250	15 Nov	1m
Pacific Coast of Aomori Pref.	2140	15 Nov	1m
Iwate Pref.	2140	15 Nov	1m
Miyagi Pref.	2140	15 Nov	1m
Fukushima Pref.	2210	15 Nov	1m
Ibaraki Pref.	2210	15 Nov	1m
Kujukuri and Sotobo Area of Chiba Pref.	2210	15 Nov	1m
Uchibo Area of Chiba Pref.	2210	15 Nov	1m
Izu Islands	2210	15 Nov	1m
Sagami Bay and Miura Peninsula	2220	15 Nov	1m
Shizuoka Pref.	2220	15 Nov	1m
Tsunami may be higher than the estimation in some places.			
Sea level may slightly fluctuate in other coastal areas but no danger.			
----- Earthquake Information -----			
Origin Time: 2015JST 15 Nov 2006			
Epicenter: 46.6North, 153.6East			
Depth: 30km			
Magnitude: 8.1			

(c) Estimated High Tide and Tsunami Arrival Time

Tsunami Information (Estimated High Tide and Tsunami Arrival Time)

Issued by the Japan Meteorological Agency (JMA)

Issued at 2030JST 15 Nov 2006

----- Estimated High Tide and Tsunami Arrival Time -----

If tsunami arrives at coasts at around high tide time, tsunami becomes higher.

Coastal Block

High Tide Time Tsunami Arrival

(continued)

Tsunami Forecasting and Warning, Table 2 (continued)

"Tsunami Warning"		
Eastern Part of Pacific Coast of Hokkaido		2110 15 Nov
Kushiro	2335 15 Nov	2130 15 Nov
Hanasaki	2338 15 Nov	2120 15 Nov
Okhotsk Sea Coast of Hokkaido		2120 15 Nov
Abashiri	2226 15 Nov	2140 15 Nov
Monbetsu	2206 15 Nov	2200 15 Nov
Esashi-ko	2228 15 Nov	2220 15 Nov
"Tsunami Advisory"		
Central Part of Pacific Coast of Hokkaido		2130 15 Nov
Urakawa	2353 15 Nov	2140 15 Nov
Tokachi-ko	2359 15 Nov	2140 15 Nov
Western Part of Pacific Coast of Hokkaido		2150 15 Nov
Muroran	2337 15 Nov	2210 15 Nov
Hakodate	0008 16 Nov	2220 15 Nov
Tomakomai-nishi-ko	0007 16 Nov	2210 15 Nov
Yoshioka	0026 16 Nov	2230 15 Nov
Northern Part of Japan Sea Coast of Hokkaido		2250 15 Nov
Wakkanai	0229 16 Nov	2310 15 Nov
Rumoi	0114 16 Nov	2250 15 Nov
Otaru	0105 16 Nov	2250 15 Nov
Pacific Coast of Aomori Pref.		2140 15 Nov
Hachinohe	0000 16 Nov	2200 15 Nov
Sekinehama	0005 16 Nov	2200 15 Nov
Iwate Pref.		2140 15 Nov
Miyako	0006 16 Nov	2150 15 Nov
Ofunato	0017 16 Nov	2150 15 Nov
Kamaishi	0011 16 Nov	2150 15 Nov
(abbreviated)	(abbreviated)	(abbreviated)
Sea level may slightly fluctuate in other coastal areas but no danger.		
----- Earthquake Information -----		
Origin Time:	2015JST 15 Nov 2006	
Epicenter:	46.6North, 153.6East	
Depth:	30km	
Magnitude:	8.1	

(continued)

Tsunami Forecasting and Warning, Table 2 (continued)**(d) Tsunami Observation Results**

Tsunami Information(Observation Results)				
Issued by the Japan Meteorological Agency (JMA)				
Issued at 2207JST 15 Nov 2006				
----- Tsunami Observation at Sea Level Stations as of 2205 15 Nov-----				
Higher tsunami may have arrived at some places other than the sea level stations.				
Tsunami may glow higher later.				
Kushiro	First Wave	2143 15 Nov	(+)	0.2m
	Maximum Wave	2155 15 Nov		0.2m
Hanasaki	First Wave	2129 15 Nov	(+)	0.4m
	Maximum Wave	2143 15 Nov		0.4m
Tokachi-ko	First Wave	2149 15 Nov		unclear
	Maximum Wave	(arriving)		
"Tsunami Warning" has been in effect for the following coastal blocks.				
Eastern Part of Pacific Coast of Hokkaido				
Okhotsk Sea Coast of Hokkaido				
"Tsunami Advisory" has been in effect for the following coastal blocks.				
Central Part of Pacific Coast of Hokkaido				
Western Part of Pacific Coast of Hokkaido				
Northern Part of Japan Sea Coast of Hokkaido				
Pacific Coast of Aomori Pref.				
Iwate Pref.				
Miyagi Pref.				
Fukushima Pref.				
Ibaraki Pref.				
Kujukuri and Sotobo Area of Chiba Pref.				
Uchibo Area of Chiba Pref.				
Izu Islands				
Sagami Bay and Miura Peninsula				
Shizuoka Pref.				
Sea level may slightly fluctuate in other coastal areas but no danger.				
----- Earthquake Information -----				
Origin Time: 2015JST 15 Nov 2006				
Epicenter: 46.6North, 153.6East				
Depth: 30km				
Magnitude: 8.1				

(continued)

Tsunami Forecasting and Warning, Table 2 (continued)**(e) Tsunami Forecast Revision**

Tsunami Warning (Revision)

Issued by the Japan Meteorological Agency (JMA)

Issued at 2330JST 15 Nov 2006

"Tsunami Warning" has been revised into "Tsunami Advisory" for the following coastal blocks.

Eastern Part of Pacific Coast of Hokkaido

Okhotsk Sea Coast of Hokkaido

"Tsunami Advisory" has been in effect for the following coastal blocks.

Eastern Part of Pacific Coast of Hokkaido

Okhotsk Sea Coast of Hokkaido

Central Part of Pacific Coast of Hokkaido

Western Part of Pacific Coast of Hokkaido

Northern Part of Japan Sea Coast of Hokkaido

Pacific Coast of Aomori Pref.

Iwate Pref.

Miyagi Pref.

Fukushima Pref.

Ibaraki Pref.

Kujukuri and Sotobo Area of Chiba Pref.

Uchibo Area of Chiba Pref.

Izu Islands

Sagami Bay and Miura Peninsula

Shizuoka Pref.

Ogasawara Islands

(f) Tsunami Forecast Cancellation

Cancellation of Tsunami Warning

Issued by the Japan Meteorological Agency (JMA)

Issued at 0130JST 16 Nov 2006

Tsunami warning has been all canceled.

"Tsunami Advisory" has been canceled for the following coastal blocks.

Eastern Part of Pacific Coast of Hokkaido

Central Part of Pacific Coast of Hokkaido

(continued)

Tsunami Forecasting and Warning, Table 2 (continued)

Western Part of Pacific Coast of Hokkaido
Pacific Coast of Aomori Pref.
Iwate Pref.
Miyagi Pref.
Fukushima Pref.
Ibaraki Pref.
Kujukuri and Sotobo Area of Chiba Pref.
Uchibo Area of Chiba Pref.
Izu Islands
Sagami Bay and Miura Peninsula
Ogasawara Islands
However, sea level is expected to fluctuate in above coastal areas.
Be careful in sea bathing and fishing.

enable recipient organizations to understand the most important information easily, namely, the necessity of evacuation. Tsunami information that follows just after the warning/advisory provides then more detailed estimates of tsunami amplitudes and arrival times for each coastal block, as well as hypocentral parameters. Examples of tsunami forecast/information are given in Table 2a–f.

Tsunami Forecast Dissemination

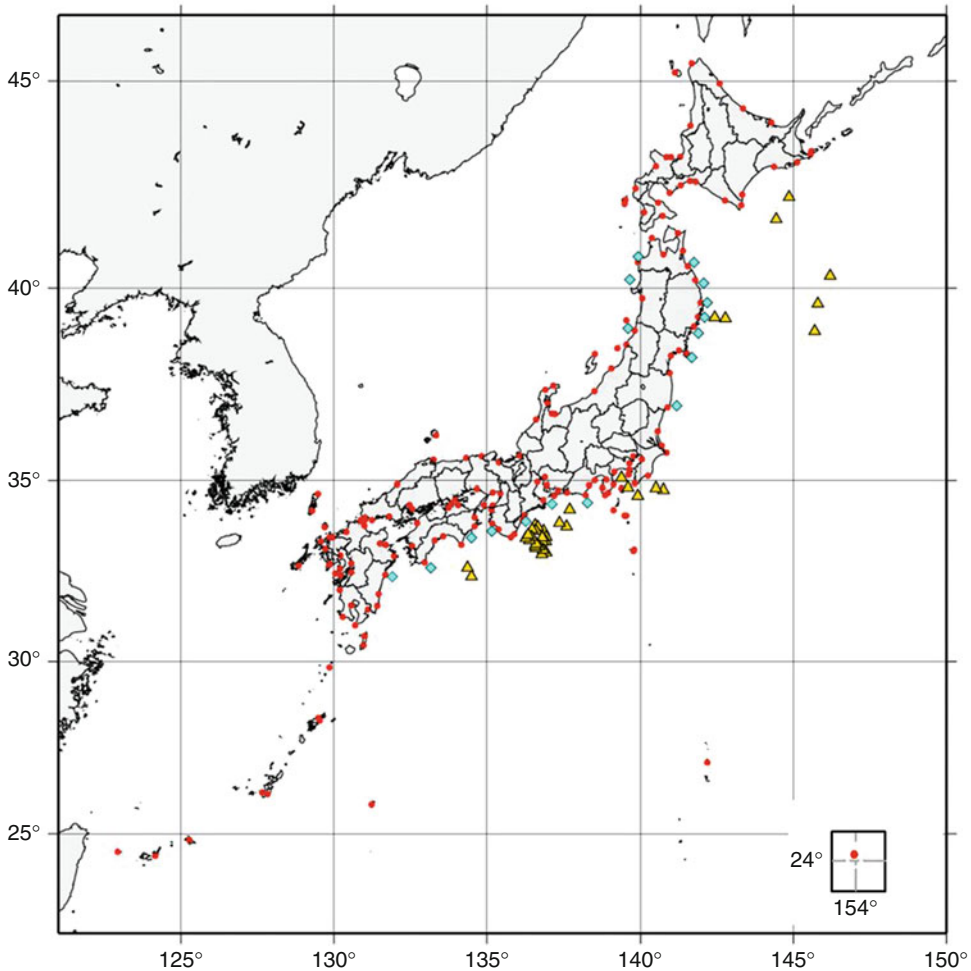
Like other forecast/information disseminated from the JMA, tsunami forecast/information is transmitted to relevant organizations by fully utilizing existing online facilities. In case of landline failure, forecast/information is transmitted through satellite link also. Loudspeaker/sirens operated by municipalities/prefectures are used in “the last mile” to reach the public. Police and fire departments have their own robust transmission routes, which are also used. Additionally, the public broadcasting company plays a very important role because of its wide outreach

and timeliness. And short e-mail services for mobile phones/terminals are spreading nationwide.

Tsunami Monitoring System

Data from about 230 (170 coastal and 60 offshore) tidal observation stations (including those operated by other institutes, like the Japan Coast Guard) are collected online at the JMA (Fig. 13). Data sampling and transmission rate is every 1 s for coastal stations, GPS tsunami meters, and cable-type (pressure sensor) tsunami meter. Those for DART buoy (pressure sensor) are 15 s sampling and 3 min transmission interval at the highest.

Arrival time, initial amplitude, polarity, maximum tsunami amplitude, and corresponding time are measured on the man-machine interface depicted in Fig. 14, after prior removal of the ocean tidal component. Tsunami observation results are issued in tsunami information. Observed tsunami amplitude is mentioned in number only after the amplitude grows larger than the criteria amplitude of one grade below



Tsunami Forecasting and Warning, Fig. 13 Tidal station network used for tsunami monitoring by the JMA. About 230 tidal stations (170 onshore and 60 offshore) are monitored in real time. *Symbols* denote

instrumentation type. *Red circle*, *yellow triangle*, and *light blue diamond* denote onshore tide gauge, ocean bottom tsunami meter (cable/buoy), and offshore GPS tsunami meter, respectively

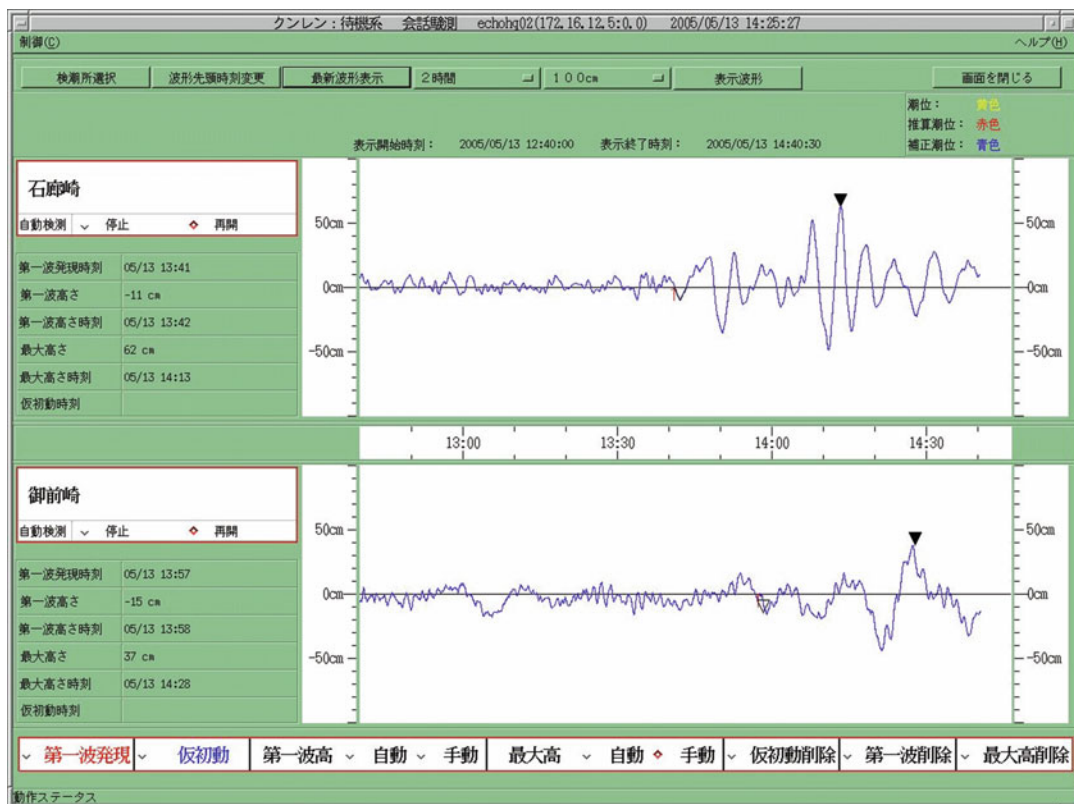
the presently valid warning or advisory (e.g., 1 m when “Major Tsunami Warning” is valid), not to give residents underestimating threat. Before then, the expression “currently observing” is used for amplitude. Initial wave arrival time and its polarity are mentioned from the very beginning, because the fact that “tsunami has arrived” is important to urge residents to evacuate.

Sea level data are used for the following: (1) revising the grade of warning/advisory in the case that estimated and observed tsunami amplitudes differ significantly by applying calibration coefficient (obs./est.), (2) informing the public on

the up-to-date status of tsunami observation, and (3) deciding about the time of tsunami warning/advisory cancellation.

Tsunami Forecast for Distant Event and Northwest Pacific Tsunami Advisory (NWPTA)

For a distant event, similar method to a local event as mentioned above is used, but the seismic waveform data from global network which are available in real time through the Internet is used for hypocenter and magnitude calculation, instead of those from domestic seismic network. The PTWC’s bulletins are also incorporated in the framework of



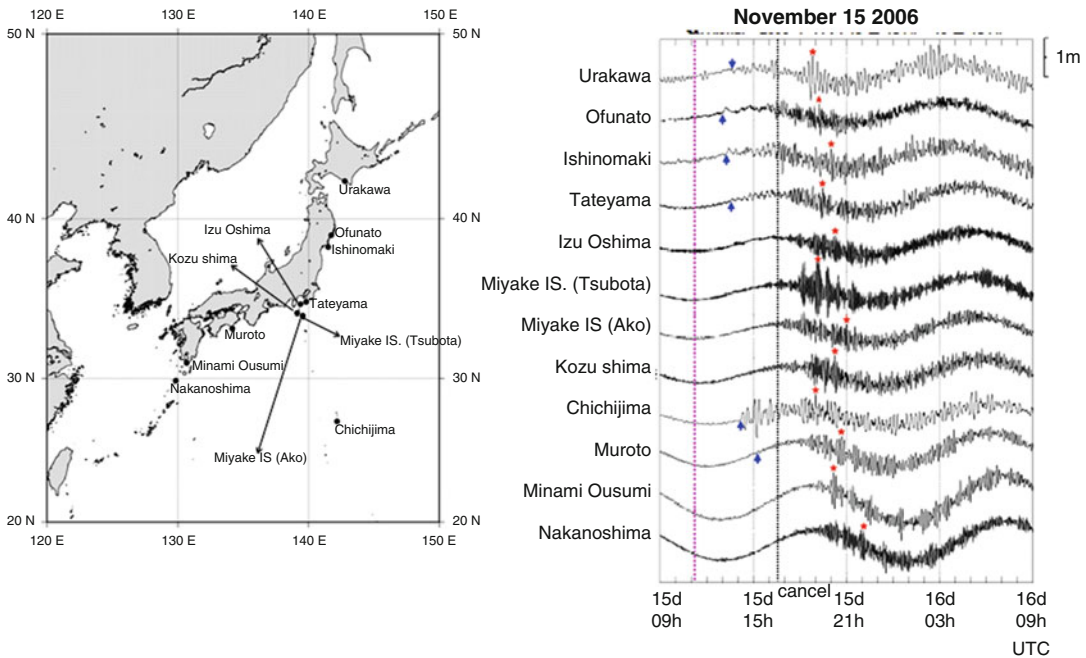
Tsunami Forecasting and Warning, Fig. 14 Example of the JMA's man-machine interface screen image in the sea level data monitoring system. Shown are two sea level recordings by different tidal gauge stations. The ocean tide component has been removed for more precise reading of

the relevant tsunami wave parameters. Arrival time, initial wave amplitude, polarity, maximum amplitude, and corresponding time are picked. The *reversed solid triangles* denote the maximum amplitude within the analyzed time window

ICG/PTWS. For magnitude estimation, Mwp (Tsuboi et al. 1995) is used commonly with PTWC and NTWC. Tsunami simulation database has been created also for distant events, taking the Coriolis force effect into account considering long-distance propagation. In case enough time is left until the earliest estimated tsunami arrival time at Japan coast, the JMA monitors sea level data which is collected via data collection platform function of geostationary meteorological satellites and circulated through GTS (Global Telecommunication System operated by the World Meteorological Organization of UN) circuit in near real time. After confirming the generation of tsunami at sea level observation sites near the source, tsunami warnings/advisories are disseminated for the same coastal blocks as for local tsunami if necessary. Estimated tsunami amplitudes in the database can

be calibrated by the actually observed tsunami amplitudes before the dissemination. Furthermore, if a reliable causing fault model is available, the JMA conducts tsunami propagation simulation for a specific setting of fault location, depth, magnitude, and mechanism so that tsunami warnings/advisories can be based on more reliable seismological analysis.

By using the same method as described above in this section, the JMA, as a regional center of the ICG/PTWS, has been providing international tsunami information (NWPTA: Northwest Pacific Tsunami Advisory) to relevant countries when an earthquake with magnitude 6.5 or larger occurs in the northwest Pacific area since March 2005. NWPTA contains estimated tsunami amplitudes and arrival times, as well as estimated hypocenter parameters.



Tsunami Forecasting and Warning, Fig. 15 Tidal station records of regular ocean tides superimposed by tsunami waves generated by the Nov. 15, 2006 Kuril Earthquake ($M_w = 8.3$). *Left*: locations of tidal gauges whose data are shown in the right figure. *Right*: tidal records of the stations shown in the left figure. Time is in UTC. The *thick blue dotted line* denotes the origin time of the earthquake, and the *blue arrows* denote the arrival

time of tsunami at each site. The *thin dotted line* denotes the time of tsunami forecast cancellation. *Red stars* mark the time of maximum amplitude at each site. At all sites maximum amplitudes were recorded about 2–5 h after the forecast had been cancelled. The largest tsunami amplitude recorded in Japan for this event was 84 cm at Miyake Is. (Tsubota)

Some Lessons Learnt from a Recent Event

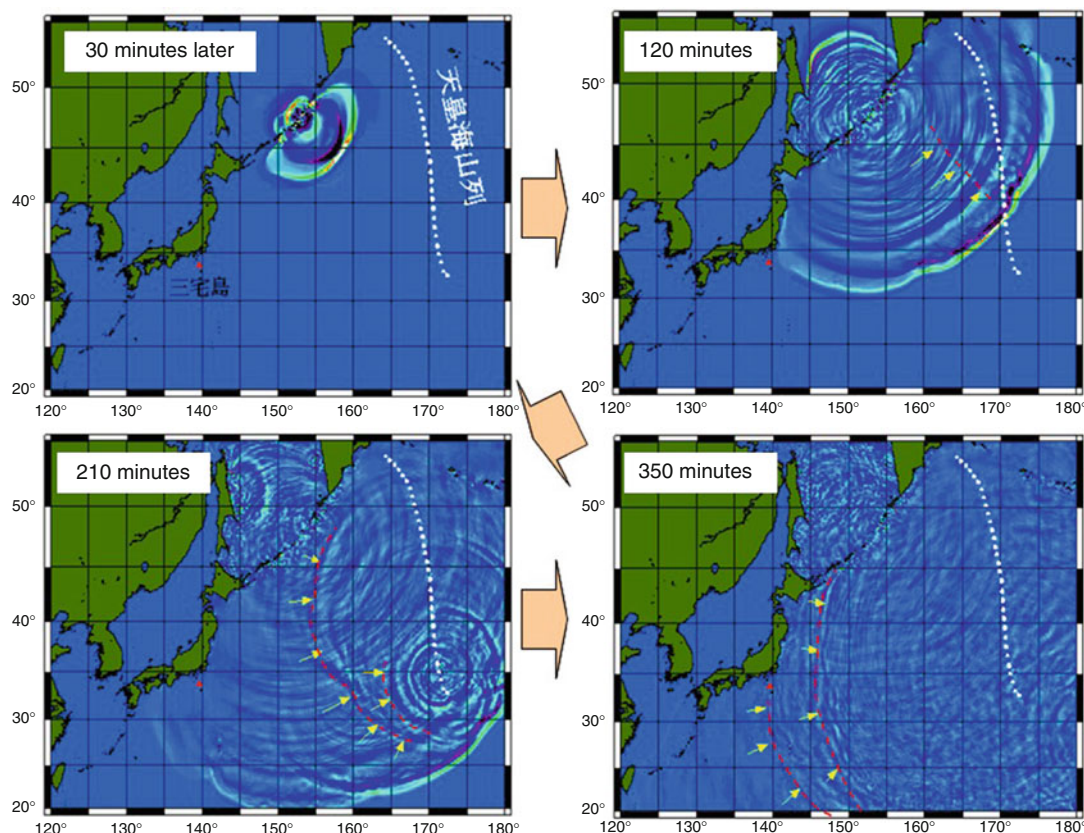
At 11:14 (UTC) on Nov. 15, 2006, a large earthquake with $M_w 8.3$ occurred in the Kuril Island Region which generated a Pacific-wide tsunami. Tsunami records at tidal gauges in Japan are shown in Fig. 15. The JMA disseminated tsunami forecasts at 11:29 (UTC) (corresponding to 20:29 JST in the related case example given in Table 2) based on the estimated tsunami amplitudes in the database. About 4 h after the first detection of the tsunami on the Japan coast, the JMA cancelled all the forecasts at 16:30 (UTC) judging from the amplitude decay in the sea level record of Chichijima station (cf. Fig. 15) that no higher tsunami will be observed. But, after the cancellation, maximum amplitudes were recorded at many of the tidal gauges along the Pacific coast of Japan up to 5 h later. The largest tsunami amplitude recorded in Japan for this event was 84 cm at Miyake Is. (Tsubota).

The JMA conducted a close examination of this case and found that these large amplitudes were tsunami waves reflected from the Emperor Seamount Chain in the middle of the Pacific (Fig. 16). Since the computational area for the tsunami simulation database of the JMA was limited around Japan (cf. Fig. 8), such reflected wave could not be represented by the simulation. Therefore, JMA examined the scenarios for which the computational area should be expanded and integrated the database.

Recent Improvements

Application of Earthquake Early Warning (EEW) Technique to Quicken Tsunami Forecast

In the near coastal area (i.e., roughly within 100 km from the coast), EEW hypocenter and magnitude estimates (Kamigaichi 2004) are equivalent to those resulting from the common procedures described above. Therefore, EEW



Tsunami Forecasting and Warning, Fig. 16 Snapshots of numerical tsunami simulation for the Nov. 15, 2006 Kuril event (Mw8.3). Coriolis force has been considered for this case. Four snapshots (30, 120, 210, and 350 min after the earthquake occurrence) are shown. The white dotted line marks the position of the Emperor

Seamount Chain. Broken red curves with arrows denote reflected tsunami waves from the Emperor Seamount Chain. The estimated arrival times of the reflected waves at the coast of Japan are consistent with the actual tidal records

results can be used as input data for tsunami forecast. JMA started this incorporation in October 2006. In the case of the Mjma6.8 Niigataken Chuetsu-Oki Earthquake of July 16, 2007, the JMA disseminated tsunami advisory 1 min after the first detection of the seismic wave.

Quick Revision/Cancellation of Tsunami Forecasts by Utilizing CMT Solutions

In JMA, automatic CMT solutions are available in 15 min or less (when the W-phase CMT solution (Kanamori and Rivera 2008) is adopted) after the earthquake occurrence. They yield more realistic magnitudes than Mjma for really great earthquakes ($M_w > 8$) or tsunami

earthquakes and reliable fault plane solutions. On the other hand, it is not proper to wait for the CMT solution for issuing the first tsunami warning, and it is not realistic neither to prepare the tsunami simulation database for so many different dip, strike, and slip angle settings with fine increments and choose the best fitting one, because it requires huge storage and computational time for the conduct of simulations for all parameter settings. Therefore, we use the CMT solution for the revision/cancellation of the first tsunami warning, and we limit the number of different fault parameter settings to 4, namely, pure reverse fault with dip angles 10, 45, and 80° and pure strike fault. Normal fault with the opposite slip angle (λ) direction to the reverse

fault with the same values for other fault parameters has almost identical tsunamigenic potential, and the difference is just a polarity of initial tsunami wave. Therefore, normal fault is treated as reverse fault with the same dip angle for this purpose. Strike angles have been fixed to the representative values of the considered regions or set parallel to the nearby trench axis or coastline. The case with the highest resemblance to the actually calculated CMT solution with respect to its tsunamigenic potential is then retrieved.

The JMA incorporates CMT solutions in the revision/cancellation of tsunami forecast as follows:

- (a) In the case of reverse or normal faults, and if the centroid depth is less than the depth determined by P and S arrival times + 30km, and Mw 0.3 or more larger than Mjma, the tsunami forecast is then upgraded in accordance with respective database cases.
- (b) In the case of strike faults, taking the database into account, and after confirmation by low or not observed tsunami amplitudes at tide gauges, the tsunami forecast is then downgraded or cancelled.

And if the first issuance is based on the preset maximum magnitude of the area (i.e., in case of possible Mjma underestimation), the first warning with significant uncertainty is revised based on the CMT solutions as well as on the observed sea level data.

Check Tools for Possible Mjma Underestimation

Although it is still difficult to accurately estimate a magnitude for gigantic or tsunami earthquake in 3 min after the earthquake occurrence, it is necessary to recognize the possibility of Mjma underestimation in about 3 min. When the possibility is detected, preset maximum magnitude of the area, instead of calculated Mjma, is adopted to avoid a tsunami amplitude underestimation in the first warning. The JMA developed several tools to realize it, and two of them are as follows:

1. Spatial extent of strong motion area (Katsumata et al. 2012)

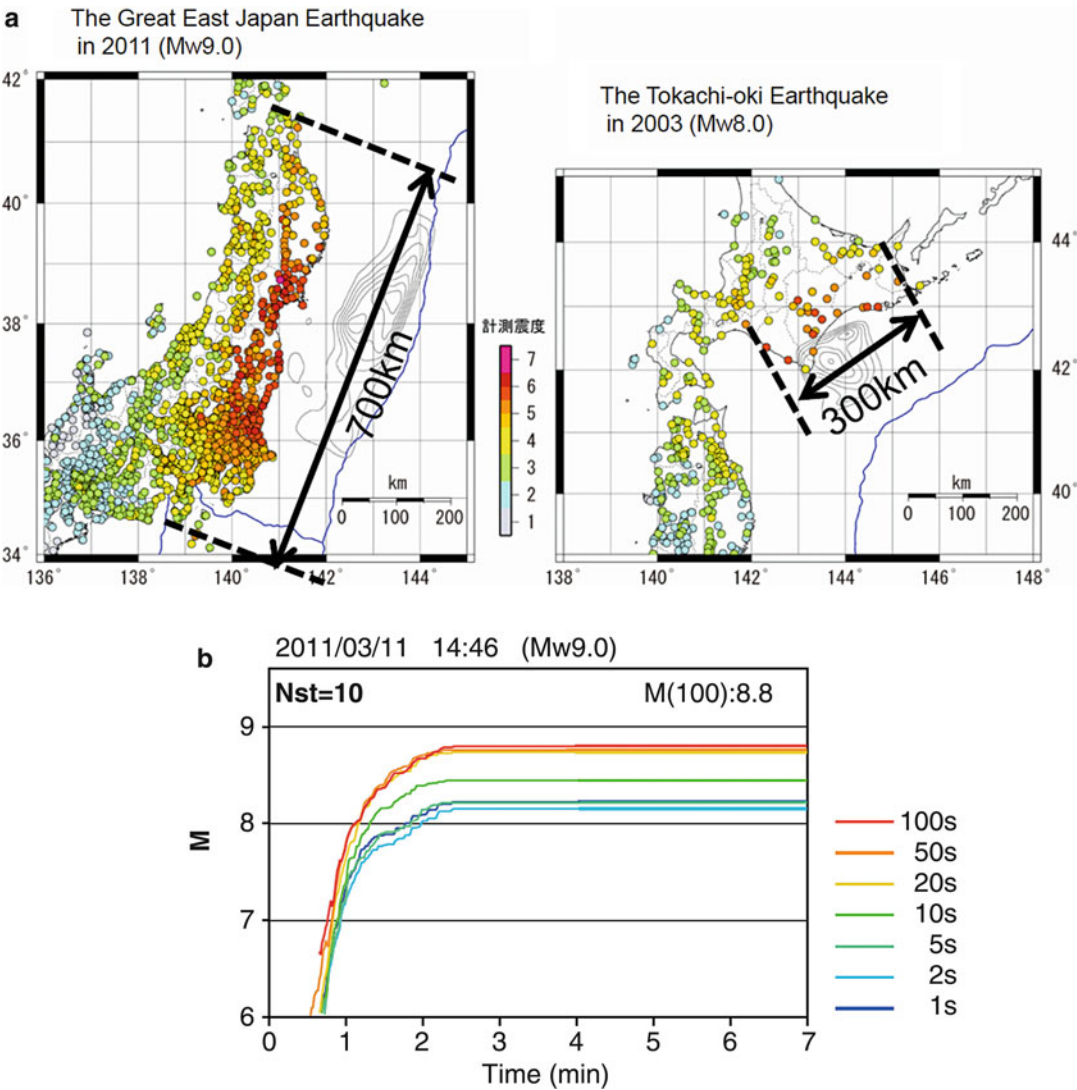
Peak amplitude of ground motion with a period of around 1 s shows clear saturation, but a spatial extent of strong ground motion area gives a good index to estimate a spatial extent of seismic fault (Mw). Measuring a span of strong motion along the trench axis gives a rough estimate on seismic fault size. Figure 17a shows the example for the 2011 Great East Japan Earthquake (Mw9.0) and the 2003 Tokachi-Oki Earthquake (Mw8.0). Quickly determined Mjma in 3 min were 7.9 and 7.8 respectively, but the span of strong ground motion area is apparently larger for the former.

2. Time evolution of newly defined magnitudes based on peak displacement amplitude with various frequency bands (Katsumata et al. 2013)

Magnitude estimation method based on peak displacement amplitude with various frequency bands (cutoff periods: 1, 2, 5, 10, 20, 50, 100 s) was developed. Figure 17b shows an example for the 2011 Great East Japan Earthquake, using closest ten station's data. From time evolution of magnitudes, rough estimate on rupture duration time can be given, and by seeing a difference of final values between long (50, 100 s) and short (1, 2, 5 s) frequency bands, the operator can recognize a possibility of Mjma underestimation.

Future Outlook

In tsunami forecasting, trade-off exists between promptness and accuracy/reliance. The JMA's tsunami forecasting strategy, especially for local events, is to satisfy both by using state-of-the-art technologies. They assure promptness of the first issuance of tsunami forecast, based on preliminary seismological results including that of EEW, and its subsequent revision, if required, as soon as more accurate and reliable data such as sea level change and the results of a more detailed complex data analysis have become available. And as for accuracy/reliability, there are two ways to be taken. One is quicken the process of reducing the uncertainty of initial tsunami wave distribution assessments. Quick focal process inversion technique, new magnitude definitions applicable



Tsunami Forecasting and Warning, Fig. 17 Examples of check tools for possible Mjma underestimation. **a** Spatial extent of strong ground motion area for the 2011 Great East Japan Earthquake (left) and the 2003 Tokachi-Oki earthquake (right). Warmer color denotes stronger ground motion. The span of ground motion larger than a certain threshold for the former is apparently longer than the latter. Thin contour curves denote slip distributions estimated from strong motion waveform data inversion analysis. **b** Time evolution of newly defined

displacement magnitudes for various frequency bands applied for the 2011 Great East Japan Earthquake case. Colors denote the cutoff periods as shown in the right. Abscissa denotes lapse time after the origin time. All curves approach to the final values after about 140 s, which is in accordance with rupture duration estimated by other studies, and final value is about 8.8 for 100 s band, whereas those for 1–5 s are about 8.2, showing a clear difference

for gigantic or tsunami earthquakes, and tsunami source inversion techniques using sea level data (also from satellite altimetry) will become effective for this purpose soon. The other way is very detailed numerical simulation, after reduction of

uncertainties in the initial tsunami wave distribution, using fine bathymetry mesh data. Along with the development of integrated calculation algorithms, the improvement of the computer performance might solve this problem in the

future. But even then, due to the stochastic nature of tsunami behavior near the coast, one has to examine carefully on what statistical quantity of simulated results the tsunami forecast criterion has to be based.

Acknowledgments We thank Dr. Peter Bormann and Dr. Kenji Satake for reviewing the manuscript, and their comments and suggestions greatly improved it.

Bibliography

Primary Literature

- Abe K (1973) Tsunami and mechanism of great earthquakes. *Phys Earth Planet Inter* 7:143–153
- Bormann P, Wylegalla K (2005) Quick estimator of the size of great earthquakes. *Eos* 86(46):464
- Bormann P, Baumbach M, Bock G, Grosser H, Choy GL, Boatwright J (2002) Seismic sources and source parameters. In: Bormann P (ed) *IASPEI new seismological observatory practice*, vol 1. GeoForschungsZentrum Potsdam, Potsdam, pp 1–94, Chap 3
- City Bureau, Ministry of Land, Infrastructure, Transport and Tourism (2011) Survey of damaged cities at the Great East Japan Earthquake, http://www.mlit.go.jp/report/press/city07_hh_000053.html (in Japanese).
- Geller RJ (1976) Scaling relations for earthquake source parameters and magnitudes. *Bull Seism Soc Am* 66:1501–1523
- González FI, Bernard EN, Meinig C, Eble M, Mofjeld HO, Stalin S (2005) The NTHMP tsunami network. *Nat Hazards* 35(1):25–39, Special Issue, U.S. National Tsunami Hazard Mitigation Program
- Hara T (2007) Measurement of duration of high-frequency energy radiation and its application to determination of magnitudes of large shallow earthquakes. *Earth Planets Space* 59:227–231
- Hatori T (1984) On the damage to houses due to tsunamis. *Bull Earthq Res Inst* 59:433–439 (in Japanese)
- Hayashi Y (2008) Extracting the 2004 Indian Ocean tsunami signals from sea surface height data observed by satellite altimetry. *J Geophys Res* 113:C01001
- Hayashi Y, Tsushima H, Hirata K, Kimura K, Maeda K (2011) Tsunami source area of the 2011 off the Pacific coast of Tohoku earthquake determined from tsunami arrival times at offshore observation stations. *Earth Planets Space* 63:809–813. <https://doi.org/10.5407/eps.2011.06.042>
- Hoshiba M, Ozaki T (2012) Earthquake early warning and tsunami warning of JMA for the 2011 off the Pacific Coast of Tohoku earthquake. *J Seismol Soc Jpn* 64(3):155–168. <https://doi.org/10.4294/zisin.64.155>, 2012, 2nd ser, in Japanese with English abstract
- Ide S, Takeo M, Yoshida Y (1996) Source process of the 1995 Kobe earthquake: determination of spatio-temporal slip distribution by Bayesian modeling. *Bull Seismol Soc Am* 87:547–566
- Imamura F (1997) IUGG/IOC TIME PROJECT numerical method of tsunami simulation with the leap-frog scheme, part 3 (Programme lists for near field tsunami), IOC manuals and guides, vol 35
- Japan Meteorological Agency (2005) A magnitude estimation using borehole volume strainmeters for earthquake events near the coast of Sumatra, Indonesia. Rep Coord Comm Earthq Prediction 74:575–577 (in Japanese)
- Japan Meteorological Agency (2013) Lessons learned from the tsunami disaster caused by the 2011 Great East Japan earthquake and improvements in JMA's tsunami warning system, [http://www.seisVol.kishou.go.jp/eq/eng/tsunami/Lessons Learned Improvements brochure.pdf](http://www.seisVol.kishou.go.jp/eq/eng/tsunami/Lessons%20Learned%20Improvements%20brochure.pdf)
- Jim Gower J (2005) Jason 1 detects the 26 December 2004 tsunami. *EOS Trans Am Geophys Union* 86(4):37–38
- Kajiura K (1970) Tsunami source, energy and the directivity of wave radiation. *Bull Earthq Res Inst (Univ of Tokyo)* 48:835–869
- Kamigaichi O (2004) JMA earthquake early warning. *J Jpn Assoc Earthq Eng* 4(Special Issue):134–137
- Kanamori H, Anderson DL (1975) Theoretical basis of some empirical relations in seismology. *Bull Seismol Soc Am* 65:1073–1095
- Kanamori H, Rivera L (2008) Source inversion of W phase: speeding up seismic tsunami warning. *Geophys J Int* 175:222–238
- Kanazawa T (2013) Japan trench earthquake and tsunami monitoring network of cable-linked 150 ocean bottom observatories and its impact to earth disaster science, Underwater Technology Symposium (UT): Conf. Proc. of 2013, 2013 I.E. Int., <https://doi.org/10.1109/UT.2013.6519911>, 5–8 Mar 2013
- Kaneda Y (2010) The advanced ocean floor real time monitoring system for mega thrust earthquakes and tsunamis -application of DONET and DONET2 data to seismological research and disaster Mitigation-MTS/IEEE OCEANS
- Kasahara M, Sasatani T (1986) Body wave analyses of strain seismograms observed at Erimo, Hokkaido, Japan. *J Fac Sci Hokkaido Univ Ser VII (Geophys)* 8:83–108
- Katsumata A (2004) Revision of the JMA displacement magnitude. *Q J Seismol* 67:1–10 (in Japanese)
- Katsumata A, Aoki S, Yoshida Y, Ueno H, Yokota T (2012) Rapid source parameter estimation of great earthquake for tsunami warning. In: *Proceedings of the international symposium on engineering lessons learned from the 2011 Great East Japan Earthquake*, March 1–4, 2012, Tokyo, pp 592–601
- Katsumata A, Ueno H, Aoki S, Yoshida Y, Barrientos S (2013) Rapid magnitude determination from peak amplitudes at local stations. *Earth Planet Space* 65:843–853
- Kawai H, Satoh M, Kawaguchi K, Seki K (2013) Characteristics of the 2011 Tohoku tsunami waveform acquired around Japan by NOWPHAS equipment.

- Coast Eng J 55(3) (2013) 1350008 (27p.), World Scientific Publishing Company and Japan Society of Civil Engineers, DOI:10.1142/S0578563413500083
- Kikuchi M, Kanamori H (1991) Inversion of complex body waves, III. Bull Seismol Soc Am 81:2335–2350
- Lomax A, Michelini A, Piatanesi A (2007) An energy-duration procedure for rapid determination of earthquake magnitude and tsunamigenic potential. Geophys J Int 170:1195–1209
- Matsumoto H, Hayashi Y, Kaneda Y (2012) Characteristics of water pressure disturbances on real-time tsunami data of ocean-bottom pressure gauges. J Jpn Soc Civil Eng Ser B2 (Coast Eng) 68(2):I_391–I_395. https://doi.org/10.2208/kaigan.68.I_391, in Japanese with English abstract
- Matsu'ura M, Hasegawa Y (1987) A maximum likelihood approach to nonlinear inversion under constraints. Phys Earth Planet Inter 47:179–187
- Nishimura T, Imakiire T, Tobita M (2010) Development on the rapid estimation for an earthquake fault model using real-time 1-second sampling GPS data. J Geospatial Inf Authority Jpn 120:63–73 (in Japanese)
- Ohta Y, Kobayashi T, Tsushima H, Miura S, Hino R, Takasu T, Fujimoto H, Iinuma T, Tachibana K, Demachi T, Sato T, Ohzono M, Umino N (2012) Quasi real-time fault model estimation for near-field tsunami forecasting based on RTK-GPS analysis: application to the 2011 Tohoku-Oki earthquake (Mw9.0). J Geophys Res 117:B02311. <https://doi.org/10.1029/2011JB008750>
- Okada M, Tanioka Y (1998) Relation of tsunami generation ratio with earthquake magnitude and hypocentral depth. Mon Kaiyo 15(special issue):18–22, in Japanese
- Okada Y (1985) Surface deformation due to shear and tensile faults in a half-space. Bull Seismol Soc Am 75:1,135–1,154
- Ozawa S (1996) Geodetic inversion for the fault model of the 1994 Shikotan earthquake. Geophys Res Lett 23(16):2009–2012
- Satake K (1989) Inversion of tsunami waveforms for the estimation of heterogeneous fault motion of large submarine earthquakes – the 1968 Tokachi-Oki and 1983 Japan Sea earthquakes. J Geophys Res 94:5,627–5,636
- Satake K (1995) Linear and nonlinear computations of the 1992 Nicaragua earthquake tsunami. PAGEOPH 144:455–470
- Shuto N (1991) Historical changes in characteristics of tsunami disasters. In: Natural disaster reduction and civil engineering. Japan Society of Civil Engineering, pp 77–86
- Shuto N (1992) Tsunami Intensity and damage. In: Tsunami engineering technical report, vol 9. Tohoku Univ., pp 101–136 (in Japanese)
- Shuto N (1998) Present state of tsunami research and defense works. Bull Coast Oceanogr 35(2):147–157 (in Japanese)
- Shuto N et al (1986) A study of numerical techniques on the tsunami propagation and run-up. Sci Tsunami Hazard 4:111–124
- Takanami T, Kitagawa G (eds) (2002) Methods and application of signal processing in seismic network operations, vol 98, Lecture Notes in Earth Science. Springer, Berlin
- Titov VV, González FI, Bernard EN, Eble MC, Mofjeld HO, Newman JC, Venturato AJ (2005) Real-time tsunami forecasting: challenges and solutions. Nat Hazards 35(1):41–58, Special Issue, U.S. National Tsunami Hazard Mitigation Program
- Tsuboi S, Abe K, Takano K, Yamanaka Y (1995) Rapid determination of Mw from broadband P waveforms. Bull Seismol Soc Am 83:606–613
- Tsushima H, Hino R, Fujimoto H, Tanioka Y, Imamura F (2009) Near-field tsunami forecasting from cabled ocean bottom pressure data. J Geophys Res 114: B06309. <https://doi.org/10.1029/2008JB005988>
- Tsushima H, Hirata K, Hayashi Y, Tanioka Y, Kimura K, Sakai S, Shinohara M, Kanazawa T, Hino R, Maeda K (2011) Near-field tsunami forecasting using offshore tsunami data from the 2011 off the Pacific coast of Tohoku earthquake. Earth Planets Space 63:821–826
- Tsushima H, Hino R, Tanioka Y, Imamura F, Fujimoto H (2012) Tsunami waveform inversion incorporating permanent seafloor deformation and its application to tsunami forecasting. J Geophys Res 117:B03311. <https://doi.org/10.1029/2011JB008877>
- Ueno H, Hatakeyama S, Aketagawa T, Funasaki J, Hamada N (2002) Improvement of hypocenter determination procedures in the Japan Meteorological Agency. Q J Seismol 65:123–134, in Japanese
- Utsu T, Shima E, Yoshii T, Yamashina K (eds) (2001) Encyclopedia of earthquakes, 2nd edn. Asakura, Tokyo, p 657, in Japanese
- Weinstein S, Okal E (2005) The mantle magnitude Mm and the slowness parameter theta: five years of real-time use in the context of tsunami warning. Bull Seismol Soc Am 85:779–799
- Wells DL, Coppersmith (1994) New empirical relationships among magnitude, rupture length, rupture width, rupture area, and surface displacement. Bull Seismol Soc Am 84(4):974–1002
- Yahagi T, Miyagawa K, Kawamoto S, Oshima K, Yamaguchi K, Muramatsu H, Ohta Y, Demachi T, Miura S, Hino R, Saida Y, Douke Y (2014) Enhancement of GEONET real-time analysis system for covering Japan. In: Proceedings of Japan Geoscience Union Meeting 2014, in Yokohama, Japan (HDS27-09)
- Yokota T, Zhou S, Mizoue M, Nakamura I (1981) An automatic measurement of arrival time of seismic waves and its application to an on-line processing system. Bull Earthq Res Inst 56:449–484, in Japanese

Books and Reviews

- Bormann P (ed) (2002) IASPEI New manual of seismological observatory practice, Vol. 1 and 2, GeoForschungsZentrum Potsdam
- Satake K (2007) Tsunamis, chap. 4.17. In: Treatise on geophysics, vol.4. Elsevier, Amsterdam, pp 483–511

Fukushima Catastrophe: The Challenge of Complexity (Collective Reflexivity, Adaptive Knowledge, Political Innovation)

Alain-Marc Rieu

Department of philosophy, University of Lyon-Jean Moulin, Lyon, France

Institute of East-Asian Studies (CNRS), Ecole Normale Supérieure de Lyon, Lyon, France

Article Outline

Fukushima: A Systemic Event, a “Black Swan”
The World After Fukushima
Bibliography

Keywords

Complexity · Catastrophe · Reflexivity · Risk ·
Epistemic disruption · Political innovation

Fukushima: A Systemic Event, a “Black Swan”

More than 3 years after the catastrophe, the initial shock is integrated in our daily lives, memories, and practices. Nothing is forgotten, but a collective and adaptive learning process is at work. From the beginning, the experience of this event opened an intense collective inquiry and learning process. This wave keeps growing. As an event, Fukushima is constantly evolving, reshaping our bodies, minds, and societies, even when we behave as if nothing had happened. In this collective learning process, the time has come to debate Fukushima’s historical meaning. This disaster cannot be reduced to an unpredictable earthquake and tsunami. Because of the size of the disaster, it is impossible to take the full measure of the event.

The world has changed. Fukushima is the name of a turning point in world history: relations between technology, politics, industry, society, and ecology are forever transformed. Its long-term impact is unpredictable and impossible to repress: wherever they live in this world, people will never see and understand nuclear energy and nuclear industries as they did before. We are learning to question the energy policies and the industries, which are exploiting our biophysical environment and producing the *milieu* in which we live. Many studies, reports, and debates have poured over the victims, the dead, the missing, and the displaced, on all Japanese, on the contaminated land and sea, on institutions, politicians, journalists, professors and experts, bureaucrats, managers, and industrial companies (Japan focus 2013). From day one, one thing was clear: the Fukushima catastrophe was an appeal for new knowledge about technology, society, democracy, and about knowledge itself. This is not an unprecedented situation: it is very similar to the Lisbon earthquake of 1755, which created debates throughout Europe and changed the conceptions of man, society, nature, god and providence, morality, science, and politics.

This entry intends to analyze how the tsunami turned into a nuclear catastrophe, which turned into an event of a level of complexity, which is still expanding. As an event, Fukushima includes also the experience, the learning process it created, its systemic impact on institutions, and established knowledge. This explains why it is impossible to think after Fukushima as we did before. This event overwhelms the institutions managing the consequences of the catastrophe and the conceptual complex in which it is interpreted. This conceptual complex associates discourses, theories, and policies gravitating around the notions of risk, trust, care, and high-tech society. To criticize this conceptual complex is to explore how to rethink *after Fukushima* the relations between technology, politics, industry, and society.

Disruptive Event

A French psychoanalyst, Jacques Lacan (1966), introduced in the 1950s a difference between reality and what is real. What is commonly called “reality” covers the objects studied and explained by science, produced by technology, and bought and used by consumers, by all of us. Reality is what makes propositions true, what common language seems to be talking about. Reality seems glued to our uses of language, to our minds, bodies, habits, and customs. Reality is the immersive constant flow of news, images, and comments in which we live, think, and communicate. Reality is what the media are showing and talking about. They build the *news* by commenting them. It is our common experience and its phenomenology, its description according to established interpretive frames.

An event such as Fukushima shows another side of reality: it shows what is *real*. What is *real* is not common reality: it is what erupts or simply happens and what punctures and deconstructs daily routines and established knowledge. It is not hidden like a secret in conspiracy theory: we just don’t know how to see it and we pretend we don’t see or understand. The real is what overflows our discourses and disciplines. Fukushima is an eruption and overflow. But the *real* is not something hidden, which suddenly appears: it has always been there in the open but unseen, unnoticed because too obvious. The Fukushima catastrophe is showing what is *real* in the case of Japan. But the shock of discovering the *real* spreads like a virus: it is a desire to find out and to know everywhere else. No country, government, or individual is immune. What is real always has a local answer, and it always is a question for everyone. To learn from Fukushima is to learn everywhere else. Fukushima does not belong to Japan alone.

What really happened in Fukushima has been expressed first by the despair and suffering of the survivors and displaced, of all victims. It has also been documented, reported, researched, and debated. The Japanese population immediately understood the catastrophe as an investigation of its causes, responsibilities, and consequences. Testimonies were a search for explanations and

explanations are also testimonies. The testimony, the reporting, and the analyses participate in the common expression and construction of *Fukushima body of experience*. This experience has been put online, partly translated into English and many other languages, made available, and debated worldwide. By becoming our common experience, the historical meaning of the Fukushima catastrophe is still under construction, and it will find its place in world history. More than three years after, the event does not fade. On contrary it remains a wave still rising in the horizon and flowing. It will flow as long as we are capable of making sense of Fukushima. If it would not flow, it would still rise like a ghost over Japan and over all of us.

This is why it is vital to continue exploring and constructing the meaning of the Fukushima catastrophe. Knowledge is the only way to emancipate from such a ghost and overcome a catastrophe. It is for Japan a historical trauma, but it is not destiny: it did not have to happen and it does not have to happen again in the future. How are the Japanese people going through this traumatic memory and experience? To pretend that life can continue as it was before is a temptation. But denial and repression are impossible: the trauma is deep and wide, a collective black hole, which is still absorbing and transforming Japanese minds. This cannot be repressed: a society cannot live long in denial. In the end, the only rational solution is to investigate and debate what has really happened, to search for the historical meaning of the catastrophe, and to accept the consequences and change. This is true in Japan, but it is also true for all those concerned in the world. This is why Fukushima is the name of a disruptive event, which has changed the course of world history by becoming a common ground for debate and research. To use a notion developed by Ulrich Beck, Fukushima is a “reflexive” event (Beck et al. 1994). Reflexivity is the collective process through which the evolution of a social system, the constraints and events changing the course of its evolution, generated the knowledge teaching individuals and groups how to criticize and reform this social system. This reflexive process characterizes modern societies and the various steps of their evolution. From this

point of view, the body of experience being built by Japanese civil society since March 2011 is questioning presuppositions, concepts, ideology, and theories at the core of all industrial societies. (This corpus includes the official report of the Fukushima nuclear accident-independent investigation commission (2012).) By doing so this transnational collective work transforms these societies and leads them to another stage of their evolution.

What Did We Learn?

How a body of experience is transformed into a body of knowledge? What sort of event is Fukushima? The answer is found in the Fukushima body of experience. The catastrophe was caused neither by the earthquake nor by the resulting tsunami: they were just the deadly trigger of a systemic catastrophe, all at once human, social, political, technological, and industrial. According to available knowledge, the *networks of power*, which decided where to build this nuclear plant and its six reactors, were the cause of the catastrophe (Nishioka 2011; Hindmarsh 2013). This power structure selected the technology; it decided the standards for the plant's construction, for its maintenance and backup systems, for the security of nearby population, and for protecting also the environment, the land, and the ocean (Crowell 2011; Koide 2011). Other nuclear plants have been built in highly seismic regions. Collective investigation established that the dangers and mistakes made were known and information was available to the media, politicians, administrators, researchers, and other experts. This comment is not an accusation but just a summary of investigation on what happened at Fukushima.

Since March 2011, collective inquiry, including various Japanese media, has uncovered the different networks power involved and their connections (Samuels 2013). What *really* happened at Fukushima is a public exhibition of the power structure controlling and managing Japanese society and economy. This power structure ("the deep state") is now naked. This power network associates various departments in the two powerful and competing ministries in charge of technological

research and energy supply, the METI (Ministry of Economy, Trade and Industry) and the MEXT (Ministry of Education, Culture, Sports, Science and Technology). It associates also the nuclear industry, utilities companies, and the industries depending on these utilities companies: the electronic and mechanical industries (mainly the car industry), chemical and metallurgical industries (including pharmaceutical and health industries), construction companies, and transportation industries. This power network associates all the industries of Japan's first and second industrial revolution, those who rebuilt Japan after 1945 and further developed Japan in the 1980s. In fact this power network owns and controls all the infrastructures of the Japanese economy and society. They are *heavy* and *hardware* industries: as such they manage the population and the territory. (These industries constitute the core of the *Keidanren*. Their power and role explain Japan's relative weakness in software, why Japan was late in developing online industries and services.) Because of their size, the range of their activities, and their accumulated wealth, these companies constitute a network of interests, which includes various factions in all political parties, which they support and finance. One proof of this is the high instability of political life caught between these power networks, social and cultural evolutions, and the geopolitical situation. (*Tenkô* (turnaround, Takeuchi 2004) is a recurrent feature in Japanese history; it is often explained and justified by personal weakness and reduced to a change of mind, when it is in fact the result of extreme pressure on individuals, groups, and even society, the result of a power struggle. "Conversion may resemble *tenkô* on the outside, but its direction is the reverse. If *tenkô* is a movement toward the outside, conversion is a movement toward the inside. Conversion takes place by preserving the self, whereas *tenkô* occurs by abandoning the self," Takeuchi (1959) 2004, p. 75.) Finally, this power network includes the media, which are largely financed and influenced by utilities companies. In summary, this power structure created what John Kingston (2012) called "Japan's nuclear village." The problem is that a closely knitted village tends to become a

“Galapagos of power” (DeWit 2012a, b). In the end, such a “village” becomes counterproductive: it generates a level of risk, which in the long-term no society can bear and no economy can afford. France is in the same situation.

This power structure emerged in the mid-1950s. By the early 1980s, it has accumulated the financial means, the expertise, and political influence to transform its aggregated power into a nationwide nuclear industry, to build nuclear plants according to their interests, to maintain these plants according to their safety and profitability standards, to distribute energy, and to manage all the population and activities depending on electricity. Nuclear energy was the perfect match for a strong and coherent power structure (Shiokura 2011). Only a power structure can decide to develop an industry supplying nearly one third of the national consumption of electricity: when all the long-term costs and risks are taken into account, no private company would consider such an investment rational.

In accordance with political theory and government studies, *sovereignty* is the name given to various technologies associating ideological constructions, political institutions, and tacit practices implemented by a group in power in order to control a population on a territory. From this point of view, nuclear energy and related industries have a sovereign function as well as the various groups associated through this technology and industries. These interest groups develop their own international relations, and through their political and administrative influence, they even have their own foreign policy. Therefore, because of its scale, complexity, and impact on the whole economy and society, its level of investment, and its intrinsic danger, nuclear power should be named a *sovereign technology*. In the world and in each country, utilities companies should also be called *sovereign industries* because they produce and distribute energy to the whole population and economy.

The problem therefore is not nuclear energy by itself but the institutional environment in which this technology is embedded, in which it was developed and is still developing. Science and Technology Studies (STS) have proven for years

that the institutional environment shapes a technology. Today, more than 2 years after the catastrophe, the nuclear industry has not renounced its objectives: nuclear energy is still promoted as the best and the more rational and economical core energy support for Japan. Since the formation of a new LDP government last January 2013, Prime Minister Abe expressed his will to restart Japan’s nuclear plants. This industry now justifies its role and explains its mission as managing the long-term transition between a fossil fuel energy system and the next, *green*, one. This strategy is similar to the European situation, in Germany, France, or Switzerland. But after Fukushima, the divide (the loss of trust) between the population, the energy-based industrial complex, and government (both the political system and the state apparatus), the people rejects this policy and strategy as well as the existing power networks behind. The problem is not that this policy is the right one or not. This divide, the public anxiety and mistrust, the shattered lives, and the contamination of land and sea are here to stay, and this anticipated “energy transition” might be constantly delayed.

For the Japanese public, according to the Fukushima body of knowledge, the real danger is the institutional environment, the power of sovereign industries controlling a sovereign technology built in the administration as a power structure. The problem we learn from the Japanese situation is all at once political, social, economic, and technological. This systemic problem is at the core of each industrial society, today in Japan but also in Western Europe (In France in particular) and tomorrow in the USA, China, Russia, and elsewhere. Japan is dramatically in advance. It is a concentrate of all the problems advanced industrial societies face today or in their future, problems also we need to understand if we want to avoid or overcome them.

Fukushima is a unique body of knowledge opening cases of investigation and debate all over the world. This body of knowledge is highly disruptive for established political systems in modern democracies. The information, explanations, justifications, and reports produced and distributed by utilities companies, by ministries, newspapers, and other media all those years

proved to be biased, partial or simply false and wrong, and sheer propaganda. People decided to believe them because they needed the jobs or simply because they were not asked (or were pressured) by central and local governments if these industries could build their plants in their neighborhoods. Secondly, we understand since March 2011 that sovereign industries are ready to compete with the government's power. *Tokyo Electric Power* (TEPCO), owner and manager of the Fukushima Daiichi nuclear power plant, has knowingly taken decisions against public safety. It is ready to defend and protect its interests against public interest and public will. As the outset of the accident, it did not inform adequately the government of the situation at Fukushima, of all dangers for the nearby population and for the power plant employees. It defied the elected government by not complying with all security and information requirements. It should be openly admitted that Fukushima is a catastrophe for modern political institutions. The catastrophe went so deep that the power structure is naked in the open, even if not yet dismantled. Beyond the tsunami, Fukushima could (it has) have happened everywhere. Some interest groups left the network in which they had collaborated because the risk for them had become too high. The press and other media chose to step with their public and took the side of the victims. Political parties were ambiguous and divided. The LPD seems in the end to stand with utilities companies and the energy industrial complex. But the former power network is not reconstituted and will probably never be. Prime Minister Abe might sound provocative by asking to restart the nuclear plants proven secure, but he does not seem ready to confront Japanese public opinion and media and public opinion. The Fukushima body of knowledge seems to have transformed the institutional system and the power structure.

Disruptive Knowledge

What do we learn from the Fukushima catastrophe? It proved forever the limitations of our present democratic institutions and conception of democracy. They do not extend to those sovereign technologies and industries, which are shaping

and managing the economy and society. (Other sovereign technologies and industries coming to mind concern the Internet (for instance, Google), banks, and other financial institutions. This problem concerns also ministries and the state apparatus itself, the so-called bureaucracy.) Contemporary democratic institutions do not touch the level of the power networks. But we have learned we always knew: democracy is a political technology designed to control social violence by transforming social conflicts into political oppositions negotiating within a specific institution, an "assembly" or parliament, with the goal to build evolving consensus and to reach decisions (rules, laws, etc.) imposed by a government on a population living in a territory within a nation state. Available political theories do not provide the concepts, problems, and methods able to analyze power networks and their impact of the long-term evolution of societies. These networks are the social infrastructures managing technological and economic structures of each industrial society. Metaphorically, power networks are the "dark matter" of social sciences as well as the "black boxes" of our societies. (Social sciences do not perceive them adequately because power networks are situated and acting in the interstices of their demarcations into various disciplines.) *Corruption* is the notion commonly used to characterize shared interests and reciprocal supports. But corruption is a moral judgment on individual members of a power network: it does not explain how a power network is organized, how it functions, and its impact at the level of the infrastructure of a nation. (There are several misconceptions. First a power network is the contrary of what "mafia" means. It is typical of advanced industries societies, distributed and not centralized, based on capital, competence, and a shared vision of a society. Power networks might generate specific risks, but they are not criminal networks. They are not hidden, but we don't fully know how to analyze them. Second to criticize the "bureaucracy" like in Japan or the "state" like in France is misleading: these two criticisms hide the right problem instead of pointing at it. I derive the notion of "power networks" from the notion of "technostructure.") Sociology studies indeed such

networks but without offering effective political innovation.

Since 11 March 2011, the *world* (all of us) learned that such ignorance is a high risk and that a disruptive event like Fukushima requires and opens new knowledge. It is still difficult to imagine what type of institutional innovation is required. But a case is made and a window is opened: in response to such destructive event, *civil society* needs to continue its collective inquiry, but it needs also to be relayed by similar *civil inquiry* in each advanced industrial nation. Knowledge outcome is already remarkable. The initial inquiry has distinguished several dimensions or layers, all connected and reinforcing each other. The first dimension concerns the *sovereign* task of government, the basic legitimacy of all institutions managing populations on a territory: they are in charge of the victims and of their needs and wishes. They are also in charge of the territory (land and sea) where this population is living or was living: their sovereign duty is to decontaminate the land, to close any source of danger for the population, and to rebuild the economic structures for these populations to develop the means to sustain their lives. When power networks control and manage society, the sovereign duty of government and state apparatus is not guaranteed to the people. The Fukushima event brings a further proof that Japan is not *really* governed by its government, but managed by power networks (van Wolferen 1989; Johnson 1995). It enforced the idea that weakness of Japanese political system is a major obstacle in time of crisis and disruption. The same observation needs to be made and proven in most advanced industrial societies: established democratic institutions are not capable to withstand the level of complexity and systemic risk of our societies.

A second dimension concerns the interaction between civil inquiry and innovative research in human and social sciences. The goal of such research is to translate this disruptive event into disruptive knowledge. The objective is to produce a body of knowledge establishing a benchmark for innovative political debates. This is my present concern. Researchers in these disciplines are mainly working in universities, but they are

closely related to intellectuals working in the media. Researchers and intellectuals are fully embedded in civil society by their students and readers. They are also directly connected to various power networks in the institutions and companies they are working for. The problem is not to take side or to support anyone as individuals but to assert the meaning and responsibility of their *work* in the situation opened by Fukushima. Their function is to explore what really happened, and this function makes them part of civil society. In this social and epistemic context, political parties would also have to respond and choose their side. It is hard to believe (but possible [In his article “Power politics: Japan’s resilient nuclear village,” Jeff Kinston (2012a, b) studies how in a typical case of *tenkô* the Noda government operated under pressure from pronuclear lobbyists a “marginalization of public opinion (. . .) in three significant policy developments”; the most significant of them was to shift the responsibility to restart nuclear plants to the Nuclear Regulatory Authority outside civil control.]) that political parties would choose to be in opposition to the majority of the population and to find themselves in an authoritarian position and even despotic situation. The media are in a similar situation: they have to choose between their patrons and their customers. To oppose this societal evolution and new role of civil society is to risk losing their legitimacy and their clients. Business communities would quickly adapt one after the other to this new context. In a fact, the knowledge generated by the Fukushima is disruptive because it has already driven our societies beyond their established conceptual and institutional frames.

In summary, Fukushima did not create a crisis. It opened a systemic transition, which is in fact already far advanced, even if still reversible. All Japan, individuals and groups, civil society, business firms and political institutions, researchers in university, and intellectuals in the media, are shifting within the former social system, transforming their relations with each other, and changing their conception of their role and this role itself. Japan is moving toward another configuration. This transition started long ago, at the end of the 1980s. (Probably in 1985 and the first

endaka (rise of the Yen, of the exchange rate of the Yen with the American dollar). It was a fatal blow to Japanese postwar competitiveness. The Japanese economy never fully recovered.) The Fukushima catastrophe has accelerated the transition, and it also gives this transition a direction and focus. For the moment this new reconfiguration seems to revolve around Japan's civil society, not only the victims of the tsunami and nuclear accident but the Japanese people who felt and understood they were betrayed by the power networks controlling their economy and society. The transition is not achieved and the direction is not set, but the tilting point is the *knowledge* produced on all these parameters.

The issue is to transform the event into knowledge and to share the knowledge of what happened throughout society. Such disruptive knowledge, effectively describing and explaining complex processes and intricate situations, cannot be reduced to established philosophical doctrines or patterns. It cannot be deduced from any transcendental idea, principle or universal values because these interpretations do not bite into power networks but simply judge from outside some of their effects. *Interpretations* are endless object of debate in the media. The knowledge extracted from the Fukushima event shows how knowledge interferes and interacts with the institutional system by analyzing processes and situations.

Far from falling into oblivion (See Takahashi Genichiro, "Resisting the force of oblivion" (*The Asahi Shimbun online*, 27 April 2013 <<http://ajw.asahi.com/article/views/column/AJ201304270009>>) and other articles concerned by a receding interest for Fukushima.), the Fukushima disaster has turned into a world event. At its core, remains the unique experience and body of knowledge constituted in Japan and still under construction. According to my assessment of the present conjuncture, in this spring 2014, the knowledge translating the experience into concepts, problems, and theories is still at an early stage of its construction. Still the problems being raised are converging toward a long-term *prospect* (not yet a program), which could already be debated:

- The first issue is to build a detailed knowledge of the power networks structuring and managing an advanced industrial society.
- The second issue is to build on this knowledge a conception of democracy fit to this level of the collective experience.
- The third issue is opening the prospect of alternative or new political institutions.
- The fourth issue is a version of the second one. The consequence of this prospect is a democratic construction of an alternative energy policy.

This prospect is not a utopia. It is borne out of what really happened in Fukushima, which means in this *world*. Nobody knows the road to reach this goal. But we have some ideas, even a method, about how to start building this road.

First, in the mid-1970s Michel Foucault introduced a new concept of *power* in political and social theory. He criticized reducing the problem of power to an expression of state sovereignty, to its institutions of control and domination. Foucault's goal was to reach beyond ideological constructions and underneath political institutions the processes at work within a society, which are shaping collective attitudes and behaviors. The political question became as follows: what makes a population governable? What makes a society? The notion of *governmentality* requires an analysis of power with the goal to study the aggregation of shared interests and local power nodes giving to this aggregate the capacity to negotiate a common policy and to implement this policy (Foucault 1976, 2004). What interested Foucault was the power to negotiate and perform a policy, to lead, to organize, and to shape individuals and groups into a manageable society.

The second problem concerns democratic theory and democratic institutions adapted to the level of power networks. It is an extension and progress toward a new democratic frontier adapted to our societies and economy driven by large-scale science and technology policies. In Japan, the project of a technological democracy has a strong history and is part of a potential response to Fukushima (Kobayashi 2012). (On Kobayashi Tadashi's method, see Callon

et al. (2001), prologue.) Kamisato Tatsuhiro (2012) tries to find in the work of Kobayashi Tadashi a method for a democratic construction of an energy policy. The debate over distributed versus centralized energy production is a conflict about the control of energy production and distribution over the economy (DeWit 2012a, b). Since January 2013, the Abe government seems to have frozen these experiments in the hope of restoring the former role of utilities companies in keeping the cost and availability of energy at the level before Fukushima. In fact the situation is stalled. The Japanese industry and society have learned to live without nuclear energy.

The method to answer these two problems is to advance without referring anymore to the pre-Fukushima situation (Iida 2012) and without referring to transcendental principles and values. There are no alternatives. It is vain to project oneself beyond the present situation: there is no other solution than to think and debate within the present situation, to continue civil investigation and collective debate, and to build disruptive knowledge. In my opinion, a method to progress is to analyze and criticize the conceptual matrices, in which the Fukushima event has been interpreted, described, and explained until now. These conceptual matrices intersect various disciplines and connect them into interdisciplinary frames. Interestingly they constitute a level deeper than established disciplines in social sciences, and these interdisciplinary frames are already a major progress in describing and explaining a disruptive event like the Fukushima catastrophe. From the beginning, the Fukushima body of experience has exceeded established social sciences. These conceptual matrices have been filters in which the Fukushima experience has been translated into a body of knowledge.

The *Fukushima body of knowledge* is structured according to four conceptual matrices: the ethics of care; a discourse on risk, one on trust; and the idea of high-tech or value-added society. These matrices acts as referentials in data mining. I will not study here the first interpretive matrix, the *ethics of care*. This conceptual matrix is clearly the most common because of its emotional intensity and practical value. It focuses on the

victims of the tsunami and nuclear contamination, on the dead, the suffering of those who survived, and of the displaced. It focuses also on the inadequate response of the central administrative and political authorities and on the solutions or lack of solutions to help the victims to cope with the disaster. The victims were and still are of course the most urgent problem. I certainly do not underestimate this frame of interpretation. But each conceptual frame responds to practical issues at a given level of the Fukushima body of experience. As a conceptual frame, the care perspective is often manipulated to put in second place other issues, to reduce a renewed and reinforced civil society to a civil society of victims.

The goal here is different. The goal is to transform a disruptive event into collective experience and to translate this experience into disruptive knowledge in order to express and construct the historical meaning of the event. The Fukushima catastrophe overwhelms the three conceptual matrices, which have been constantly projected on the event and which are reducing its level of complexity. These interdisciplinary matrices intersecting existing social sciences are all we have at our disposal to grasp the full measure of this event. The only solution is to enter inside these three matrices and analyze their presuppositions in order to extract the concepts and problems leading to epistemic innovation. It is a typical Schumpeterian process of creative destruction. Clearly these four matrices emerged and were conceptualized well before Fukushima, in the 1980s and early 1990s, including Michel Foucault's concept of power.

For all these reasons, what *really* happened at Fukushima Daiichi is not an internal Japanese affair. A powerful conceptual complex replicated in all advanced industrial nations was torn apart like the buildings covering the nuclear reactors. This complex manages, explains, and justifies the interactions between research, innovation, industry, government, and society. These three layers are closely connected with each other. The Fukushima body of experience has opened the possibility to analyze each of them and how they are connected. These three matrices are deeply embedded in our societies, and each of them has

been since the 1980s the subject of an influential book, which opened until today debates and research in politics and the media and social sciences. The goal is to criticize each conceptual matrix and in turn this complex in order to open a collective debate and research on the resulting situation.

Risk as Metaphysics

The theory of risk is foremost a theory to evaluate and manage the probability of risk in a cost/benefit analysis based on the notion of assessment, precaution, and prevention (Godard et al. 2002). But it is also much more than this today. The idea of risk has become metaphysics, a conception of society, and even vision of contemporary life in society. To mention Lacan's distinction between reality and "what is *real*" has for goal to avoid and criticize metaphysical interpretations of the Fukushima catastrophe referring to an overpowering nature, uncontrollable by mankind and unpredictable by human technology, whatever precautions (seawalls, dykes, backup and security systems, evacuation policies and exercises) human endeavors can imagine and build. This metaphysical interpretation is the very metaphysics TEPCO is developing since the catastrophe. TEPCO asserts that it did all it could, before and after, because a tsunami of this magnitude could not have been reasonably predicted. These ideas are trivial: nature cannot be controlled by mankind, the wish to control nature is vain and even dangerous because nature is always more powerful than any human anticipation and precaution, etc. Accidents of this magnitude are supposed to be fate: the duty of humans is to edify their own world and nature is always there to dwarf in the end human ambition.

This popular metaphysics has a long history. It was best expressed by Martin Heidegger's conception of technology (Heidegger 1977). In 1950, after Nazism, the Second World War, the Holocaust and Hiroshima, the German philosopher was denouncing the folly of modern humanity of "enframing" (For definition of *enframing*, its connection to technology and modern science, see Heidegger 1977 pp. 20–23.) (*Gestell*) nature, to dominate and control nature. Nature was

supposed to have been objectified by "modern" science and then reduced by technology to resources to exploit, accumulate, and distribute. For Heidegger, the source of this folly was western metaphysics, which he contrasted with ancient Greece's conception of humanity living in harmony with nature. Heidegger's philosophy had and still has a great influence: the Fukushima catastrophe seems to prove him right. In his essay on technology, Heidegger gives as an example a dam on the river Rhine: it *enframes* not only the water flow but the myths and imagination of a whole German culture and society. ("What the river is now, namely, a water power supplier, derives from out of the essence of the power station," (pp. 16–17).) Fukushima Daiichi nuclear plant was from the beginning of the catastrophe a perfect substitute for the dam on the Rhine. This metaphysics has two main drawbacks. Utilities companies share the same philosophy: nature is overwhelming and cannot be *enframed* without high cost. It dissimulates the real question, which does not concern the power of nature but the power of utilities companies on society and the economy. Any Heideggerian discourse on the Fukushima catastrophe is ineffective.

An advanced expression of this metaphysics is found in Ulrich Beck's influential book, *Risikogesellschaft* (1986) (*The Risk Society*). The German sociologist developed an alternative and updated perspective on the problems raised by Heidegger and his commentators. The strength of the book is to propose a broad conception of risk, from ecological and industrial risk to social and individual risk, including new forms of subjectivity. The notion of risk becomes a picture and a vision of the various evolutions reshaping industrial societies since the 1980s. In this perspective, risk cannot be reduced to the precautionary principle and to risk analysis. Deeper than an ideology, it is a metaphysics proving that all industrial societies have entered a "new modernity." Beck's goal is to conceptualize from a sociological point of view this new "modern social order" (It is also a response and refutation of postmodern cultural studies (Beck et al. 1994) strongly influenced by Heidegger's disciples.) (part 3). The source of these transformations is the ecological transition,

which erupted in the 1970s with the first massive energy crisis. Today, especially since 2006, this energy crisis has proven its full depth and massive impact: it has altered the conditions of economic and social development of all industrial societies. This evolution justified and is still used to justify nuclear energy.

From the beginning, the Fukushima activated Beck's conceptual frame, in Japan and abroad. The problem with a conceptual frame is that it filters the event and constructs reality. It also determines and even prescribes a range of comments on this reality. Events generate in social systems a reflexive process, which is structured by existing conceptual frames. This frame is a major progress: the energy and ecological crisis and challenge have replaced the metaphysics of nature. This crisis has driven industrial societies beyond the historical opposition between external risks (natural disasters) and internal risks proper to *techno-industrial societies*, whether "old," "mature," or "newly industrialized." Ulrich Beck explained in the 1980s that this opposition between external and internal risks had vanished: it had become an obstacle to a proper understanding of the present social, economical, and ecological condition of industrial societies. For Beck, this opposition could not be restored: this means that one cannot expect ecology to define a model or a norm, which industrial societies should comply with. Natural risks are internalized within contemporary societies. They cannot be interpreted as natural or external accidents: social and economic systems are responsible for most natural disasters. (From flooding due to deforestation and anarchic urbanization to global warming due to carbon emissions by industries and unsustainable energy consumption and due to transport and urban, suburban, and even rural lifestyles.) Social and economic systems are themselves the source of various natural catastrophes, from the Chernobyl to Fukushima, from industrial pollution and overexploitation of lands to climate change.

The overcoming of the opposition between nature and society, between two distinct and conflicting orders, (In the European context, this taboo opposition dates from ancient Greece.) is

breaching a deep anthropological order in all industrialized societies. The idea of "risk society" expresses this transgression and the Fukushima catastrophe condenses collective fear and anxiety. It is a transgression, not an anthropological transgression but an institutional transgression by an aggregated network of interests. The problem is to evaluate if this conceptual frame describes adequately the situation and provides the means to formulate problems that can be researched and debated in a democratic society. The metaphysics carried by the conceptual frame is an obstacle to analyze an event like Fukushima, to extract knowledge from a dramatic collective experience. What was called "nature" in the past and still in our imagination has become the ecology of social and economic systems, and this ecology is transforming the biophysical conditions of life on the planet. For Ulrich Beck, this diagnosis is the proof of a "new modernity": all social and economic problems are ecological problems, the reverse being also true. They have become one multidimensional process, which has extended the level of risk and disequilibrium beyond institutional control and political management. The fear is that this disequilibrium has initiated an irreversible and uncontrollable evolution, beyond the extent of the precautionary principle. (The precautionary principle is a risk management technology. Its goal is not to reduce risk but to assess the limits within which a project is financially rational for a company or a state, justified and tolerable by a population or a government, which is supposed to represent and protect this population.) This is whole debate on climate change. Fukushima is a multidimensional catastrophe. What is feared is the uncertainty of the situation: uncertainty is a call for new knowledge. (For a distinction between risk and uncertainty, see Callon et al. (2001), pp. 37–55.) A metaphysical approach increases fear and anxiety: it expresses the complexity of the experience without extracting knowledge from Fukushima body of experience.

A critical point is reached. Beyond the conceptual frame, in a "risk society," the real issue is the knowledge of the networks of power organizing and managing a society, exploiting the biophysical environment within an energy policy. This

knowledge is disruptive because it gives a new meaning to politics and to the relevance and structure of the political process. This is the present step of the reflexive process initiated by the Fukushima catastrophe. This is a problem we all have in common. This proves that knowledge, its institutions, methods, and practices constitute today the core of our societies. It proves also that research and democracy are closely intertwined. In their influential book drawing on several experiments in Japan and Europe, Callon et al. (2001) explain that the time has come to imagine and organize democratic institutions adapted to a “risk society.” A “technological democracy” has for goal to manage risk according to collective rational decisions within a democratic process. Beyond rhetoric on participatory democracy (Hindmarsh 2013), this is exactly what Kobayashi Tadashi (2012) has been experimenting in response to the Fukushima catastrophe.

The Paradox of Trust

In my opinion, for the moment, the most important lesson from Fukushima is that shared knowledge is the common ground for democracy in an advanced industrial society. In order to place this role, knowledge needs to be extracted from existing conceptual frames. Once again, existing conceptual frames are not false or wrong, but they construct and restrict the meaning of disruptive events. Since March 2011, discourses on risk have been closely related to discourses on trust. Trust in government and institutions had been breached, trust was the basis of the social bond, and trust had to be restored. Trust is a powerful conceptual frame with many different aspects and discourses. The problem is to evaluate how it fits the Fukushima experience, if it produces adequate knowledge of this experience.

Indeed a “risk society” depends on trust in the future, in institutions, in knowledge, in experts, in entrepreneurs, and above all in rational expectations. Trust is an endless quest, always missing and in search of an impossible certainty. Like risk in Ulrich Beck’s book, the return of the question of “trust” marks a historical moment expressed in an influential book written by Francis Fukuyama (1995). Sign of those neoliberal times, in the

introduction, Fukuyama declares it to be a “book on economics,” not in political theory. In a few years, trust would become a worldwide field of research and debate. After having predicted the “end of history” when the Soviet Union had collapsed and China turned to a market economy, F. Fukuyama was now trying to explain the emergence and consequences of the neoliberal movement. Since the 1980s, neoliberalism was revolutionizing, one after the other, all industrial societies, including former communist nations as well as Japan. As we still remember, it was supposed to open an age of worldwide economic growth and political freedom. F. Fukuyama’s intention was to explain that neoliberalism could not be reduced to a set of economic techniques to be learned and applied around the globe. It was based on culture and values, on those “social virtues” best exemplified in American history: a strong work ethics (p. 45) and a “spontaneous sociability” (p. 46) had created “an art of association” between hard work and knowledge in successful entrepreneurial projects. Trust was supposed to be the key explanation of the spirit of American capitalism. But this spirit was cultural, deeper than history, customs, and institutions. But this culture of capitalism was not an ethnic feature proper to a given people or civilization: it could spread to other nations, be adapted and adopted. Neoliberal policies had also a strong impact on Japanese politics and policies since the early 2000s, especially under Prime Minister Koizumi: the goal was “to loosen” Japan’s institutional system.

Fukuyama’s argument relies on a distinction between low-trust and high-trust societies. This distinction modifies the usual opposition between traditional and modern (capitalist) societies, between community and society. Strong family values create high trust within each family, clans, or people. But they also intensify low-trust between families, clans, and people. According to Fukuyama, this explains why nations based on family, clans, and ethnic groups are usually poor, underdeveloped, and under the authoritarian power of one group on all others. It hinders overall economic and social progress because any change would alter the balance of power between families

or ethnic groups. In these conditions, a common economic and public sphere cannot emerge. By comparison, a high-trust society, like the USA, England, Germany, and Japan, is a type of society where clannish ties and ethnic bonds have evolved toward growing individualization and toward a strong public sphere. Families and ethnic communities did not disappear, but individuals are in charge of their own destiny and of the well-being of their families. To achieve these modern goals, individuals need to cooperate with each other and to build association and common enterprise in a public space emancipated from ethnic, religious, or clannish control. Histories are many, but the spirit common to all liberal societies is based on “sustaining sociability” between large groups of people. This is the ideal version of the “melting pot,” the model of a political community of individuals. This conceptual frame is extremely influential in Japan: traditionally Japan understood itself as an extended family unified by the emperor and a strong sense of nation. This ethnic and sacred conception of trust could not withstand the history of the Japanese people in the twentieth century. But a different version is quite meaningful since the 1945 and is even reinforced since Fukushima. High trust within the Japanese people is combined with low trust between Japanese civil society and its government and bureaucracy. This low level of trust is considered hindering Japan’s recovery from the systemic crisis, which started in the early 1990s and from reconstruction since March 2011.

As a vision of society, neoliberalism is simply a rearrangement of the relations and hierarchies between society, government, the economy, and religion. It established during the 1990s the conditions of the globalization process as well as the conditions of the systemic crisis, which started in 2007. This vision and ideology had a major influence on all industrial nations. This increased differentiation had in each nation a similar effect: it introduced a void and a gap at the core of society. It made societies more *flexible*. But the various functions constituting a social system do not fit anymore into a coherent whole. Economic interests, models, and values have taken precedence. People in their daily life, facing personal and

private problems, have developed the feeling that they are left outside the dynamics of society. They feel the need to reorganize and make sense of their lives at another level. Paradoxically neoliberalism has introduced a collective and individual experience based on distrust and anxiety. The resulting situation is quite different from Fukuyama’s diagnosis in the 1990s. In retrospect, neoliberal policies have created a long-term *divide* between the economic sphere and the *social* sphere or “civil society.” In fact the new degree of autonomy granted to the economy led slowly to a new degree of autonomy granted to society, to individuals and groups in this society. This type of autonomy, which has been taking shape since the 1980s, is quite different from both modern political freedom and the moral (sexual) and social emancipation of the 1960s.

This emptying and reconfiguration of society had three main consequences. As first consequence, the problem of ethics and law found a new urgency and became a major political issue. Typically, according to John Rawls’ minimalist concept of society, what people have in common in contemporary societies was reduced to a conception of justice and fairness, to a subjectively acceptable degree of equality or inequality. The second consequence was mainly seen in the USA, as well as in other regions and countries. This emptying of society and the resulting personal anxieties generated a wide return to religion and spirituality in the USA and in the rest of the world. Religion became a last refuge for many of those who were losing their jobs and social identity.

The third evolution is the most important today because it leads neoliberalism to a new stage and possibly its progressive overcoming. It is particularly influential in Japan. This evolution is the reinvention of the modern ideal of “civil society” and of the capacity of individuals and groups to find between themselves, at their own level and in their immediate communities, and the capacity and the will to develop solidarities, common projects, and forms of resistance, mainly against government and businesses. In the end, neoliberalism is not all economic: it is also characterized everywhere by strong social movements for the environment and for food safety and against pollution

and nuclear energy, against GMOs, against new fields of medical research (like stem cells), against tyranny and corruption, and against also the present systemic crisis. There are strong interactions and confusion between these various expressions of “civil society,” religion or spirituality, ethics, and social movements. This return of “civil society” is highly significant even if the themes associating people mainly concern their immediate lives and interests. What is significant is strong commitment and active resistance (For instance, “les insoumis” in France, “Occupy Wall Street,” in the USA, in Spain, Greece, Italy, Russia, etc.). This surge and reinvention of civil society might be the most enduring consequence of neoliberalism. A similar evolution happened in Japan since the 1990s. It is intensified by the Fukushima nuclear catastrophe and resulting collective inquiry to a point where the whole institutional system is in question.

The question of trust is a major issue when put in its proper context. But what brings back trust once it is broken? Since the 1980s, trust has increasingly become a major requirement in business, research, and politics, superseding traditional respect for social elites and even for competence. Trust is a subjective and shared certainty. It must be repeatedly proven and this proof remains a psychological experience according to which individuals judge politicians, professors, experts, companies, and their managers. Neoliberal values and policies might strengthen for a given moment the economy, but the social basis of a neoliberal economy remains unstable and its political legitimacy basically weak and constantly questioned because society is reduced to an experience of trust, on ethical values, and it is judged according to benefits to individuals and groups. In the case of Japan, the long-term crisis, the political failures, and finally the Fukushima catastrophe have destroyed trust in institutions. But it has also reinforced the need and hope for “civil society,” for interactions between individuals and groups outside and against the present power structure. This conjuncture is opening in Japan a collective need and a search for a renewed democratic progress. If this search for democracy is once more betrayed, Japan will indefinitely lose

the coherence and collective resilience required to achieve the transition, which started in the early 1990s. The reference to trust can only be achieved by reforming the political system, uprooting it from existing power networks and by turning to civil society. This is exactly what the Fukushima body of experience and body of knowledge have been doing since 11 March 2011. Trust is always what is missing. But after Fukushima, mistrust turned into a collective feeling of betrayal and injustice, of a final divide between civil society and government, the bureaucracy, and big companies. In order to make sense of this collective feeling and transform it into a constructive political movement, the shared knowledge of the situation needs to become a political platform for debate, research, and reform. The challenge is to operate a transition from a value-based civil society to a knowledge-based civil society.

The High-Tech, Value-Added, or Knowledge-Based Society Model

This is the third conceptual matrix: it is far more influential and embedded in the network of influence and power governing Japan since the late 1970s. It is both an ideology and a program adopted and adapted by all industrial nations. It is extremely powerful and it largely escapes inquiry because the discourses and ideologies expressing the two other conceptual matrices are not to be touched. Once again, this matrix is not false or wrong: it is just there, an essential part of the epistemic infrastructure of all advanced industrial societies. It determines the conception of competitiveness and productivity. It explains how nuclear energy became the core of this type of society. It is the conceptual source of our science and technology policies, and it explains why they turned into research and innovation policies in the 1980s. This third matrix articulates the relations between science, technology, politics, and the economy. Society is considered as the milieu in which these interactions take place. Since the 1980s, research and innovation policies are supposed to drive industrial societies beyond their present economic and social problems. Today they are supposed to bring solutions to ecological challenges and to provide innovations

capable of restoring competitiveness and creating new industries. These industries are supposed to create jobs paying for new taxes, which will finance social programs as well as these science and technology policies in an endless spiral of innovation and progress. This ideal vision and its related policies have different names: the formation of a “high-added value economy” or “new economy,” more generally “knowledge society” (Rieu 2005). These theories, discourses, and policies have their source in mid-1980s Japan. Sakaiya Taichi (1985) published in 1985 *The Knowledge-Value Revolution or a History of the Future* in which he explained that “the zenith of the industrial society” has been reached: the “postwar petroleum culture” ended in the 1970s with the energy crisis.

For him a new technological revolution has started: it generates “a shift in demand that will favor the consumption of knowledge value rather than natural resources” (introduction). The added value extracted from research and innovation has always in the past and would continue in the future to supersede the value of natural resources. Today this idea is considered so obvious that it is not even questioned anymore. It is a remarkable book, but in retrospect with a problem: nuclear energy is never mentioned, even if nuclear fusion is (p. 146) as well as solar energy. The reason is simple: “the end product is the same electric power produced by hydroelectric or oil-or-coal-burning plants.” So how electricity is produced does not matter, only cost and availability matter. Why is nuclear energy “under the radar” of Sakaiya Taichi, himself, a MITI official? Nuclear energy is not “under the radar” but put “under the carpet” in order to avoid any conflict with antinuclear movements. But there is another reason: beyond huge cost and high risk, nuclear energy is typically a high-value-added technology. In the imagination of the time, it does not depend on vast amount of natural resources, complex transportation, and transformation processes. It is supposed to depend on small quantities of *rough* uranium, technologically enriched in order to become combustible in a nuclear plant. According to this reasoning, beyond the scale and complexity of nuclear energy processes, it is assumed to be a

dematerialization of energy, the triumph of technology over matter. Sakaiya Taichi knew that the very scale and complexity of nuclear energy production led to the formation of immensely powerful nuclear energy utilities complex. He could not foresee that this power network would in the end be responsible for the Fukushima disaster. Finally, the rhetoric of “knowledge over matter” hides that the development of nuclear energy production in Japan took off in the early 1980, as a response to the first petroleum crisis. Before Fukushima, nuclear power plants were producing only 25 % of Japan’s electricity (against 73 % in France). But this quantity is enough for utilities companies to regulate the availability and cost of electricity to industry and society. This explains the argument that nuclear plants have to be restarted. It also explains the role played by nuclear energy in a high-tech economy. The Fukushima catastrophe is the end result of a complex ideological, political, technological, and economic construction. Criticism of this conceptual complex opens the way for a better understanding of the post-Fukushima situation.

Also at stake are the idea of high-value-added economy and the project of a knowledge-based society. We now understand that in the late 1970s and early 1980s, nuclear energy was considered the most efficient response not only to the energy crisis but also to the idea that research and development (R&D) was the proper Japanese answer to the rising cost of energy and all natural resources. This was in early 1980s a major change in the conception and organization of Japan’s industrial policy: R&D was considered the core of all present and future industrial policies and the basis of future economic and social progress. R&D became the object of policies of increasing ambition all along the 1980s. When the bubble burst at the end of the 1980s and early 1990s, R&D policies were considered the only answer to Japan’s economic crisis and the only solution for its future. If industrial policies found in nuclear energy the solution to secure energy, R&D policies were concentrating on many different sources of energy, including solar, thermal, coal, wind, etc. In summary, nuclear energy technology was at the core of industrial-nuclear complex, but it

was not the core of the various science and technology policies launched since the 1990s. They are two connected but also different networks of power, with different interests and values. It is an important moment: in Japan, research and innovation might finally become the backbone of economic and social development. It is a major shift in the power structure managing Japan's long-term evolution. The research and innovation field of activity is taking the lead in Japan's future evolution, with important political consequences.

This shift cannot be achieved without the support of Japanese civil society. Since 2008, before the catastrophe, the trust of the population in science and technology was considered a major issue. Studies have discovered that the population had a high trust in scientists and engineers but a low trust in scientific and research institutions because of their close links with the power networks. The Fukushima catastrophe just confirmed the suspicion of the Japanese. The research and innovation communities consider quite important to connect with "civil society" in order to explore new objectives and possibilities. Because science and technology are considered today the basis of long-term social and economic development, because these policies are more and more inclusive and involve all aspects of life in society, research and innovation policies have learned since the mid-2000s to take into consideration any strong public opposition and civil resistance. Research policies of this scale cannot be implemented against public opinion and civil society. Since 2006 and the 3rd Basic Plan, Japanese science and technology policies explicitly refer to public opinion and social needs. In fact, for the last 15 years, public debates are organized in a rigorous and even exemplary fashion (Callon et al. (2001), prologue.) (ScienceWise 2011). Arimoto Tateo (2006) explained how these procedures were embedded in the conception of the 4th Basic Plan for science and technology. (Former director of the *Research Institute of Science and Technology for Society* (RISTEX) in the Science and Technology Agency at the prime minister office.) Since March 2011, a tension and even a divorce are obviously running deep between sovereign industries and society but also

between these industries and those research institutions, which since the 1990s are orienting Japan in a different direction. Indeed these policies heavily finance and promote research on new energy sources, energy, and transport. Still to associate the population in the design of these large-scale policies remains a difficult theoretical and practical challenge.

The World After Fukushima

My goal in this entry was to explain that the Fukushima catastrophe is still happening. Its historical and world meaning is not yet established. It cannot be reduced to the earthquake, the tsunami, and the nuclear catastrophe and its consequences on the population and the environment. It is a systemic event, all at once geological, environmental, technological, social, economic, and political. But it is also an epistemic catastrophe because it defies our conceptual capacity to grasp the extent and depth of the catastrophe. The problem is as follows: how a society, its institutions, its established knowledge, its scientific and engineering disciplines and expertise, and its social sciences can respond to such a high level of complexity? The challenge of such an event is the information it generates in a population within a social system. Since 11 March 2011, the response of the Japanese people has been to start a collective inquiry on the catastrophe. Everyone participated, by their testimonies of anger or despair and by their study, inquiry, and debate on all aspects of the catastrophe. The Internet has been collecting all this information and transforming it into a collective experience.

But another problem followed immediately: how does a collective experience become capable of explaining and of establishing responsibilities leading to reforms? By transforming this experience into shared knowledge. The paradox is that knowledge is a social and historical construction, but a systemic catastrophe like Fukushima exceeds established disciplines and even interdisciplinary conceptual frames. It also exceeds political institutions. The problem becomes as follows: where such a catastrophe is leading a nation?

Where does it drive a social system? How to make sense of such complexity? The solution is to accept the challenge of complexity and to explore as far as possible this new situation. I don't pretend to provide solutions. I just try to extract from the collective response of Japanese society a platform for new research and debate. These research and debates touch the core of our societies and question the capacity of our institutional system to respond to complexity.

Bibliography

- Arimoto T (2006) Innovation policy for Japan in a new era. In: Whittaker H, Cole R (eds) *Recovering from success: innovation and technology management in Japan*. Oxford University Press, Oxford, pp 237–254
- Arimoto T (2011) Japan's new science and innovation policy in a changing world. In: Conference of Japan Science and Technology Agency, Washington, DC. 12 Jan 2011
- Beck U (1986) *Risikogesellschaft*. Suhrkamp, Frankfurt
- Beck U, Giddens A, Lash S (1994) *Reflexive modernization. Politics, tradition and aesthetics in the modern social order*. Stanford University Press, Stanford
- Callon M, Lascoumes P, Barthe Y (2001) *Agir dans un monde incertain. Essai sur la démocratie technique*. Seuil/MIT Press, Paris/Cambridge
- Crowell T (2011) The roots of Fukushima. *Asia Sentinel*, 24 March
- DeWit A (2012a) Japan: building a Galapagos of power? *The Asia-Pacific Journal* 10(47):3
- DeWit A (2012b) Distributed power and incentives in post-Fukushima Japan. *The Asia-Pacific Journal* 10(49):2
- DeWit A, Iida T (2011) The 'power elite' and environmental-energy policy in Japan. *The Asia-Pacific Journal* 9(4):4
- Foucault M (1976) *La volonté de savoir*. Gallimard, Paris
- Foucault M (2004) *Naissance de la biopolitique*. Gallimard, Paris
- Fukuyama F (1995) *Trust. The social virtues and the creation of prosperity*. Simon & Schuster, New York
- Godard O, Henry C, Lagadec P, Michel-Kerjan E (2002) *Traité des nouveaux risques*. Gallimard "Folio", Paris
- Heidegger M (1977) The question of technology (1950). In: *The question of technology and other essays* (trans: W. Lovitt). Harper and Row, New York
- Hindmarsh R (2013) *Nuclear disaster at Fukushima Daiichi*. Routledge, London
- Iida T (2012) What is required for a new society and politics: the potential of Japanese civil society. *The Asia-Pacific Journal* 10(46):1
- Japan Focus (2013) <http://japanfocus.org/Japans-3.11-Earthquake-Tsunami-Atomic-Meltdown>
- Johnson C (1995) *Japan: who governs?* Norton, New York
- Kamisato T (2012) How to formulate a strategy for energy supply. *Nippon.com*, Sept 11
- Kingston J (2012a) Japan nuclear village. *The Asia-Pacific Journal* 10(37):1
- Kingston J (2012b) Power politics: Japan's resilient nuclear village. *The Asia-Pacific Journal* 10(43):1
- Kobayashi T (2012) The reality of Japan's *National Debate*: Citizen participation and the nuclear energy issue. *Asteion (Hankyu-Communications)* 77: 192–208
- Koide H (2011) The truth about nuclear power: Japanese nuclear engineer calls for abolition. *The Asia-Pacific Journal* 9(31):5
- Lacan J (1966) *Le séminaire sur la lettre volée (1956)*. In: *Ecrits, Le seuil*, Paris, pp 11–61
- Nishioka N (2011) Toward a peaceful society without nuclear energy: understanding the power structures behind the 3.11 Fukushima nuclear disaster. *The Asia-Pacific Journal* 9(52):2
- Official report of the Fukushima nuclear accident independent investigation commission (2012) *The National Diet of Japan*. http://naaic.go.jp/wp-content/uploads/2012/07/NAIIC_report_lo_res.pdf
- Rawls J (1972) *A theory of justice*. Harvard University Press, Cambridge
- Rawls J (1993) *Political liberalism*. Columbia University Press, New York
- Rieu A-M (1996) Japan as a techno-scientific society: the new role of research & development. National Institute for Research Advancement Review, Tokyo
- Rieu A-M (2005) What is knowledge society?, STS Nexus, Santa Clara University, Center for Science, Technology and Society, San Jose, Sept. <http://halshs.archives-ouvertes.fr/halshs-00552293/fr/>
- Rieu A-M (2011) Beyond neo-liberalism: research policies and society. The case of Japan. *Copenhagen J Asian Stud* 29(2):58–78
- Rieu A-M (2012) *Penser après Fukushima. Technologie souveraine et contrôle démocratique*, Tokyo, revue *Ebisu* no 47, Spring-Summer 2012, pp 69–78. <http://hal.archives-ouvertes.fr/hal-00701743>
- Rieu A-M (2013) Japanese translation Sonoyama Chiaki. Akashi Shoten, Tokyo, Avril
- Sakaiya T (1985) *The knowledge-value revolution or the history of the future*. PHP kenkyujo, Kyoto. English translation (1991) Kodansha International Tokyo
- Samuels R (2013) *Disaster and change in Japan*. Cornell University Press, Ithaca
- ScienceWise (2011) International comparison of public dialogue. <http://www.sciencewise-erc.org.uk/cms/sciencewise-erc-resource-library/>>
- Shiokura Y (2011) Asahi Shimbun, 18 Aug 2011. Comment un pays irradié est devenu pronucléaire, translation *Courrier international*, Paris, 23 Aug
- Takeuchi Y (2004). What is modernity? (1959). In: Calichman R (ed) *What is modernity*. Columbia University Press, New York, p 75
- Van Wolferen K (1989) *The enigma of Japanese power*. Norton, New York

Alain-Marc Rieu is professor of contemporary philosophy and science studies at the University of Lyon-Jean Moulin and senior researcher at the Institute of East-Asian Study at the Ecole Normale Supérieure de Lyon. He is studying the mutation since the 1970s of the role, conception, and

organization of knowledge in advanced industrial societies. He now concentrates on a comparative study of science and technology policies from the perspective of their institutional environment (<http://am-rieu.name>).

Part II

Volcanic Processes, Eruptions, and Hazards

Volcano Seismology: An Introduction

Vyacheslav M. Zobin
Observatorio Vulcanológico, Universidad de Colima, Colima, Mexico

Article Outline

Glossary
Definition of the Subject
Introduction
Seismic Signals Generating During the Magma Migration Within the Earth's Crust
Seismic Signals Generating During the Movement of Magma Within Volcanic Conduits
Seismic Signals Associated with the Surface Manifestations of Volcanic Activity
Relationship of Volcanic Earthquakes with Eruption Process
Future Directions
Bibliography

Glossary

Earthquake The rupturing along the fractures and faults as well as the eruptive activity produces the seismic waves, which could vibrate the Earth surface as *earthquakes*. The site of initial radiation of seismic waves in the interior is *earthquake focus* or *hypocenter*; its projection to the surface is *epicenter*. The seismic waves are divided into *body waves* and *surface waves*. The body waves *P* (compressional waves with volumetric changes) and *S* (shear waves without volume change) radiate from the earthquake focus; the surface *Rayleigh* and *Love* waves are formed as a result of the interaction of body waves with the surface of the Earth. The record of seismic waves is a *seismogram*. The earthquake size is described by *magnitude M* (or *m*), the conventional value proportional to the

earthquake energy, and by *seismic moment Mo*. The *focal mechanism* characterizes the stress system acting in the earthquake source zone, while the *stress drop* gives the value of stress release during the earthquake. Earthquakes may form the *foreshock-aftershock sequences* with a main shock and the *swarm* sequences without any main shock.

Volcano A *volcano* is a site at which molten rock, known as *magma* and formed in the depths of the crust of the Earth, reaches the surface of the Earth from the interior. The material ejected through the vent of volcano frequently accumulates around the opening, building up a *volcanic edifice*. Between the main elements of a volcano, there are a *volcanic cone* that is the result of the accumulation of ejected material around the vent and a *crater* that is the surface connection of the *volcanic conduit* through which the ejected material reaches the surface. The crater may be situated on the summit of the cone (*summit crater*) or on a slope of the cone (*flank crater*). Beneath a volcano, we have the magma feeding system of the volcano that consists of a deep *magma reservoir* or *chamber* which is a reservoir of magma in the shallow part of lithosphere from which volcanic material is derived; an intermediate *magma storage*, a place between the magma chamber and the Earth's surface, where magma may be collected before an eruption; and the *feeding conduit*, connecting the magma collectors with the surface. The surface appearance of magma is called *lava*.

Volcanic eruption A *volcanic eruption* is a process of material transport from the Earth's depths to the surface during the activity of volcanoes fed by *basaltic* (containing <52% SiO₂), *andesitic* (containing 52–66% SiO₂), or *dacitic* (containing >66% SiO₂) magmas. Volcanic eruptions may be of *central* (or summit) type with an eruption at the summit crater; *flank* (or lateral) type, with an eruption on the flank crater; or *fissure* type, when the eruption takes place from an elongated fissure, rather than from

a central vent. Explosive eruptions are commonly classified as follows: *Strombolian* explosions, producing eruption columns of tens or hundreds of meters in height; *Vulcanian* explosions, which are characterized by a small amount ($\leq 1 \text{ km}^3$) of erupted products and higher eruption columns (reaching 10–20 km altitude); and *Plinian* explosions whose eruption columns may reach up to 45 km altitude.

Definition of the Subject

Volcano Seismology is a science about seismic signals originating from volcanoes and associated with volcanic activity. Keiiti Aki, the outstanding seismologist of the twentieth century, wrote: “The subject of *Volcanic Seismology* is not only the most beautiful and spectacular, but also the most difficult to study of all the subjects seismologists have encountered on Earth. This is because seismic sources in volcanoes involve dynamic motion of gas, fluid and solid, and propagation paths in volcanoes are usually extremely heterogeneous, anisotropic and absorptive, with irregular topographies and interfaces including cracks of all scales and orientations. Thus, volcanic seismology is the most challenging to seismologists requiring ingenuity in designing experiments and interpreting observations” (Aki 1992, p.3). The study of the origin of these signals generated by volcanic activity, their spatial-temporal distributions, and their relationship with volcanic processes and using them as a tool to investigate the deep structure beneath volcanoes and in attempts to predict a volcanic eruption together create the subject of *Volcano Seismology* that lies in the interaction between volcanic and seismotectonic processes.

Introduction

Volcano Seismology has its roots in the ancient times. Pliny the Younger, who gave us the first known scientific description of the eruption of Vesuvius in 79 A.D., wrote about numerous earthquakes related to this eruption. Vesuvius occupies a special place in the history of the study of seismic signals associated with volcanic activity. It was the first volcano whose earthquakes were mentioned in scientific literature, it

was the first volcano to have a volcano observatory in 1847, and it was the first to be monitored using seismological equipment. The Palmieri electromagnetic seismograph, built in 1862, was the first seismic instrument to record volcanic seismicity. The seismograms were written on a telegraph tape. Later, in 1903, the Bosch-Omori seismograph was installed at Mont Pelée, Martinique, 1 year after the great eruption of 1902 (Ferrucci 1995). Seismic observations at Kilauea, Hawaii, volcano had begun in 1912 by H. Wood just after the foundation of Hawaiian Volcano Observatory (Wood 1913).

Strictly speaking, the science of *Volcanic Seismology* was born when the Japanese seismologist Fusakichi Omori began his systematic investigations of seismic signals related to the 1910 eruptions of Usu and Asama volcanoes (Omori 1911–1913, 1912) and the 1914 eruption of Sakurajima volcano (Omori 1914, 1922). Omori defined the volcanic earthquake as “seismic disturbance, which is due to the direct action of the volcanic force, or one whose origin lies under, or in the immediate vicinity of, a volcano, whether active, dormant, or extinct” (Omori 1912, p. 8).

A three-component seismic station, with a magnification of 100 and the period of the pendulum about 4 s, was installed by Omori in July–August of 1910 at a distance of about 1–2 km from the craterlets of Usu. This station allowed him to record the earthquakes originating from the volcano and the volcanic micro-tremors whose period was about 0.5 s. The micro-tremors were observed only during volcanic activity. Omori demonstrated a good relation between the appearance of micro-tremors and explosions of the volcano.

Omori also noted that many eruptions of Japanese volcanoes were preceded by numerous earthquakes. “In such a case, he wrote, seismograph observations made at the vicinity of the center of volcanic activity would give people a warning of the approaching outburst” (Omori 1911–1913, p. 38). Omori’s analysis of volcanic earthquakes recorded in the course of volcanic activity had shown the great potential of seismic observations.

The basic investigations performed later at Asama Volcano Observatory in Japan, at Kamchatka (Klyuchi) Volcano Observatory in Russia, and at Hawaii Volcano Observatory in the USA had become the foundation for the construction of

Volcano Seismology as a formal science. The installing of broadband digital equipment at volcanoes during recent decades gives a good perspective for better understanding the nature of seismic signals originating with volcanic activity (Chouet 1996).

The seismic signals originating from volcanic activity have waveforms that are different from those of tectonic earthquakes. Studying the seismic signals during the 1910 Usu eruption, Omori (1911–1913, p. 36) wrote about the “...occurrence besides the proper volcanic earthquakes, of well defined small quick unfelt vibrations, which may be termed micro-tremors.” The first classification of volcanic earthquakes, which is still commonly employed, was proposed by Minakami (1960). He divided them into four types according to the location of their foci, their relationship to the eruptions, and the nature of the earthquake motion: deep (type A) and shallow (type B) volcanic earthquakes, volcanic tremor, and explosion earthquakes. At other volcanoes, seismologists have developed their own classification with more detailed descriptions of every subtype of earthquake (Tokarev 1966; Latter 1979; Malone 1983; Lahr et al. 1994; among others). Figure 1 shows broadband seismic records illustrating a great variety of volcanic earthquakes that may be recorded during a volcanic eruption (Zobin 2017).

The main types of the seismic signals associated with volcanic activity can be divided into three broad groups depending on their relationship with eruption process: the signals generating during the magma migration within the Earth’s crust, the signals generating within the volcanic conduits, and the signals generating during the subsurface or surface manifestations of volcanic activity. In this entry, I discuss the waveforms and the nature of the seismic signals of these three groups and their relationship with volcanic activity.

Seismic Signals Generating During the Magma Migration Within the Earth’s Crust

Geological Processes Occurring During Magma Migration

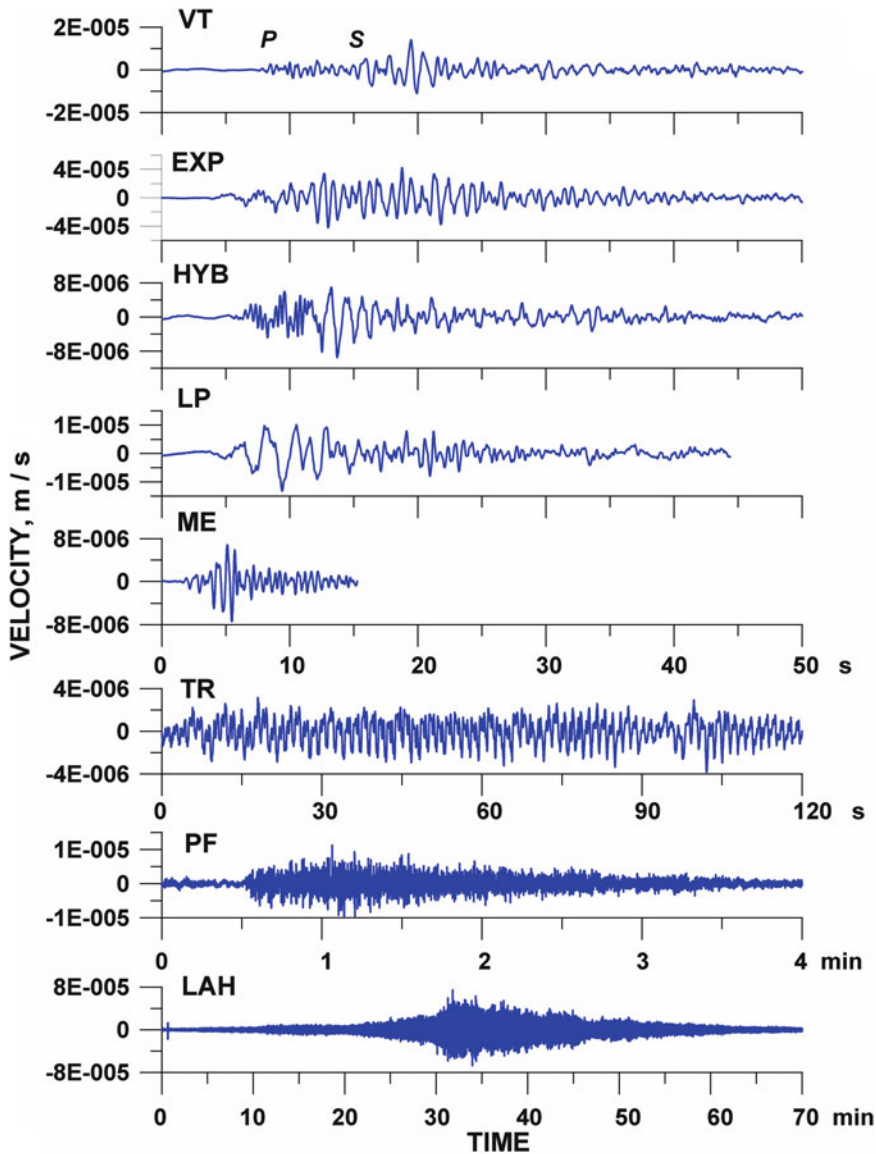
Molten rock (magma) forms in the upper part of the mantle and then rises through the ruptured

crust to the surface along zones of ocean spreading and subduction. Partial melting of rocks within the mantle generates basaltic magmas, whereas generation of silicic magmas generally also involves partial melting of crustal rocks (Rogers and Hawkesworth 2000). After generation, magma commences its journey to the surface. When magma rises from its storage zone in the crust, it begins to move along the system of pre-existing or newly formed fractures and the volcanic conduit to the surface as a dyke or sill. The magma ascent rate is a function of the pressure in the magma storage region, the density and viscosity of the magma, the conduit diameter, and the resistance to flow in the conduit that connects the magma storage zone to the surface (Rutherford and Gardner 2000). All these factors affect the generation of the volcano-tectonic seismic activity beneath the volcanoes and define the style of eruptive processes.

Volcano-tectonic earthquakes are caused by shear or tensile fracturing during magma movement from depth to the Earth’s surface through conduits and dykes. These earthquakes occur within and around the volcanic edifice and reflect the interaction of two general geological processes: the magma migration to the Earth’s surface and the crustal tectonic activity. The precise location of small (magnitude 1–2) volcano-tectonic earthquakes that were observed during the 1983 dyke intrusion fissure eruption of Kilauea volcano in Hawaii (Rubin et al. 1998) illustrates the effects of interaction of magmatic and tectonic processes in generation of volcano-tectonic seismicity along the east rift zone (Fig. 2a). The foci of earthquakes are located about 0.5 km south of the eruptive fissure, but a dyke, that dipped only 7° from vertical, would pass through them. The hypocenters of the earthquakes lie between 3 and 4 km depth (Fig. 2b), indicating the origin of seismicity within a very narrow depth interval.

Waveforms and Spectra

Figure 3 shows the seismograms of volcano-tectonic earthquakes that were recorded at Usu volcano, Hokkaido. The waveforms of volcano-tectonic earthquakes (VT, see Fig. 1) are



Volcano Seismology: An Introduction, Fig. 1 Types of seismic signals (velocity, vertical component) recorded during the 1998–2011 unrest at andesitic Volcán de Colima. *VT* volcano-tectonic earthquakes, *EXP* seismic signals associated with magmatic explosive events, *HYB* hybrid signals associated mainly with white-cloud degassing, *LP* long-period signal recorded at different stages of unrest, *ME* microearthquake signals, forming together with short hybrid signals the microearthquake

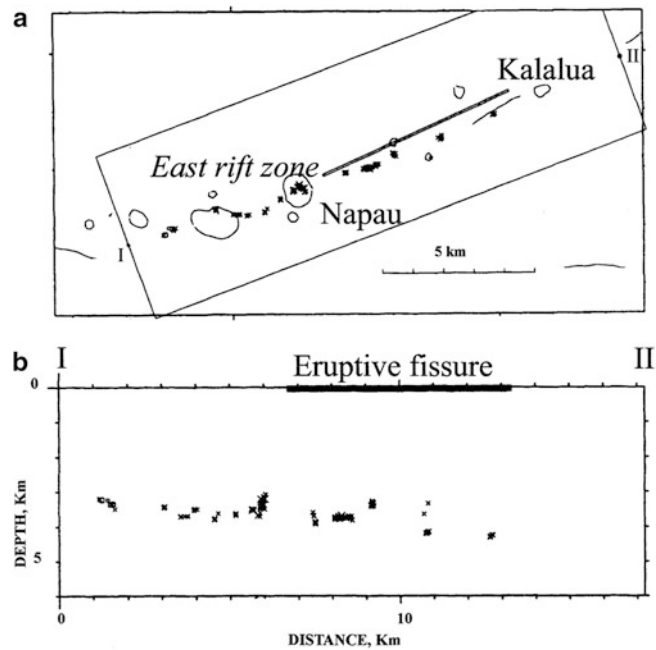
swarms, associated with extrusive and explosive activity, *TR* volcanic tremor recorded at different stages of unrest, *PF* signals associated with pyroclastic flows or rockfalls of incandescent rocks, and *LAH* signals associated with lahars. All seismic signals were recorded by broadband seismic station at a distance of 4 km from the crater and corrected for instrument response. (Courtesy of Colima Volcano Observatory)

characterized by clear arrivals of *P*- and *S*-waves. The spectra of *P*- and *S*-waves have the dominant frequencies between 1 and 20 Hz (Fig. 3c);

S-waves are characterized by lower-frequency content comparing with *P*-waves. These waveforms are similar to those of tectonic earthquakes.

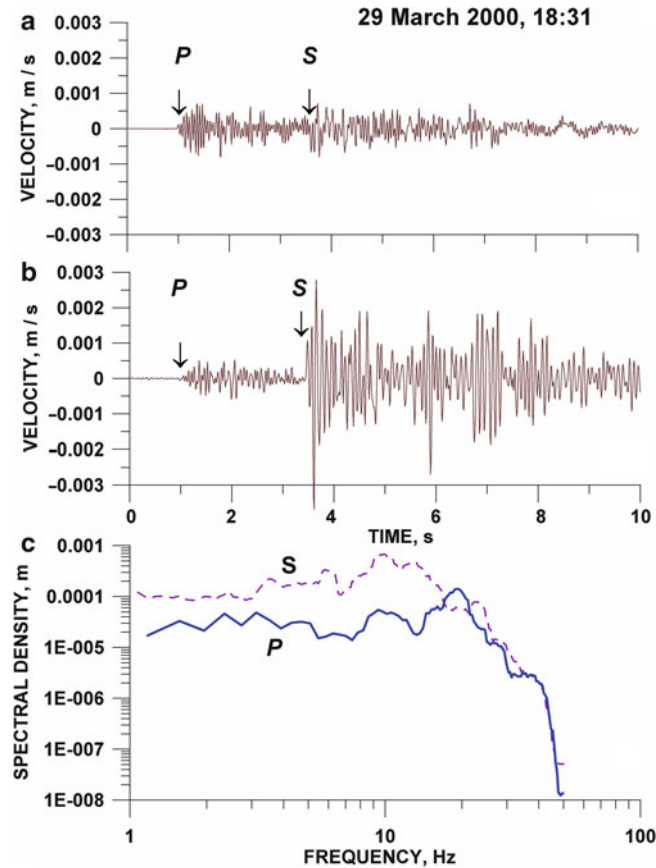
Volcano Seismology: An Introduction,

Fig. 2 Clustering of volcano-tectonic earthquakes (crosses) along the eruptive fissure (thick line) during the 1983 dyke intrusion in east rift zone of Kilauea. (a) Epicenters of earthquakes; (b) cross section along the line of eruptive fissure. The cinder cones are shown by ovals. (From Rubin et al. 1998)



Volcano Seismology: An Introduction,

Fig. 3 Waveforms of vertical component (a) and horizontal N-S component (b) of seismic signal and Fourier spectra of *P*- and *S*-waves (c) of volcano-tectonic earthquake (velocity, corrected for instrument response) preceding the 31 March 2000 flank eruption at Usu volcano, Hokkaido, and recorded at a distance of about 20 km by a short-period station. (Courtesy of Japan Meteorological Agency, Sapporo, Japan)



Source Modeling

The source of volcano-tectonic earthquakes, produced by movement along tectonic fractures, may be represented by a deviatoric part of the seismic moment tensor, which is responsible for production of fracture. The deviatoric part may be decomposed into two source types (Fig. 4a): a double couple (DC) and a compensated linear vector dipole (CLVD). The dominant type may be discriminated by the relationship between the smallest and largest eigenvalues of the seismic moment tensor ε ; it is equal to 0 for a DC source and to 0.5 for a CLVD source.

Seismological practice and theoretical studies showed the double couple source (Fig. 4b) as the most adequate for shear faulting that produces volcano-tectonic earthquakes. The values of ε that were obtained from the Harvard CMT solutions for large volcano-tectonic earthquakes, related to the eruptions of some volcanoes of the world, varied from 0.10 to 0.18, indicating DC sources (Zobin 1992). The model predicts the quadrant distribution of *P*-wave first motions using stereographic projection and allows the discrimination of the three main types of faulting in earthquake foci: strike-slip, thrust, or normal fault.

The non-double couple sources, or the sources with values $0.2 \leq \varepsilon \leq 0.5$, were determined for a few volcano-tectonic earthquakes located at ring fault structures in Iceland (Nettles and Ekström 1998) and at Etna and Vesuvius, Italy (Saraó et al. 2001). Shuler et al. (2013a) identified 101 earthquakes with vertical CLVD focal mechanisms that took place near active volcanoes all over the world between 1976 and 2009. The focal mechanism solution for the non-double couple sources may be obtained by modeling with synthetic seismograms; it has a form shown for CLVD in Fig. 4a.

Shuler et al. (2013b) found that ring-faulting mechanisms explain many characteristics of vertical-CLVD earthquakes, including their seismic radiation patterns, source durations, association with volcanoes in specific geodynamic environments, and the timing of the earthquakes relative to volcanic activity. At the same time, they concluded that physical mechanisms related to fluid flow and volumetric changes are

incompatible with seismological, geological, and geodetic observations of vertical-CLVD earthquakes.

Seismic Signals Generating During the Movement of Magma Within Volcanic Conduits

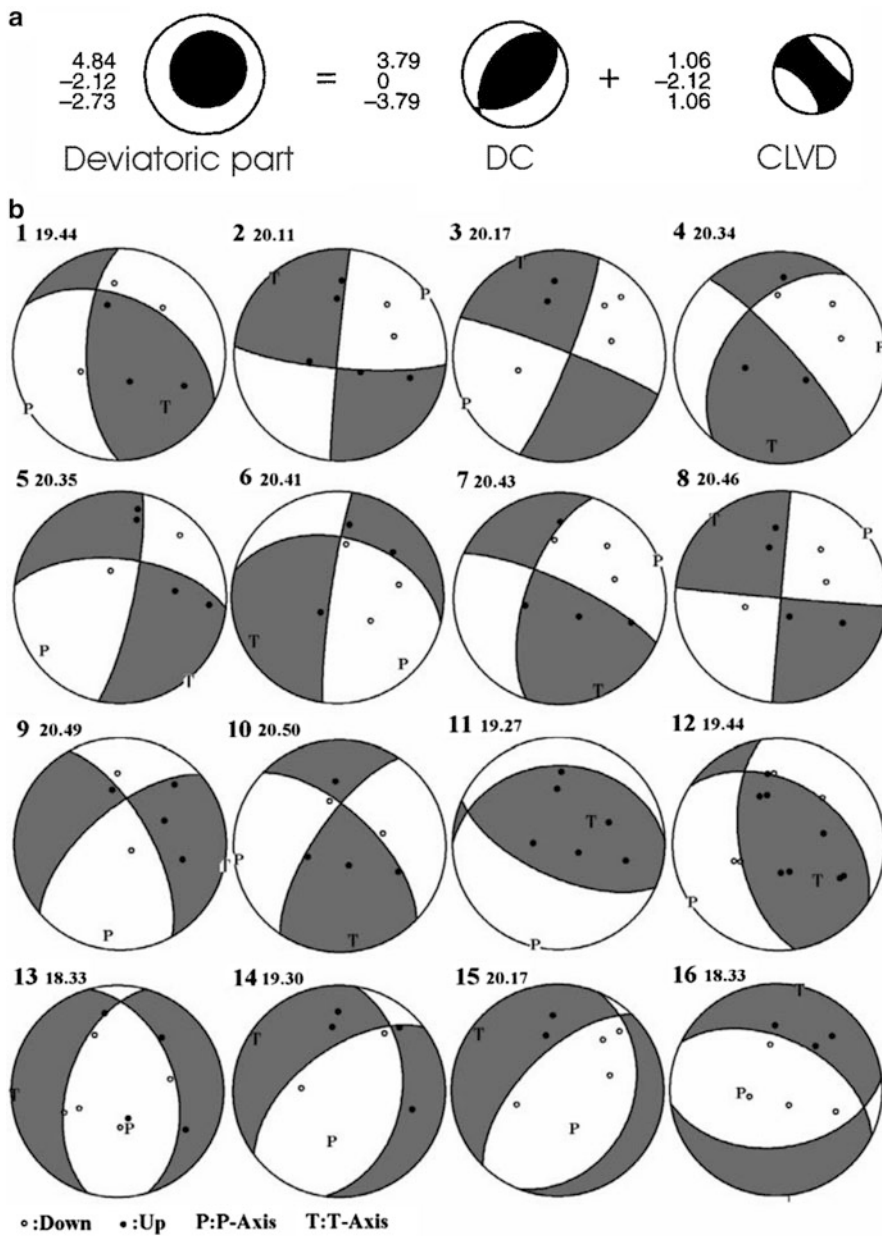
Geological Processes Occurring During the Movement of Magma Within Volcanic Conduits

During its ascent in the upper crust within the volcanic conduit (Fig. 5), the magma undergoes exsolution and fragmentation, thus, changing from a continuous, homogenous melt with a dispersed gas phase to a two-phase mixture of melt fragments within a continuous, gas phase (Cashman et al. 2000). The seismic signals, generating at this stage of magma ascent to the surface, reflect the dynamic interactions between gas, liquid, and solid along geometrically complex magma transport paths (Chouet and Matoza 2013). Networks of cracks can be formed within the magma body. The cracks, filled by fluid phases, are supposed to be excited into resonance by a pressure transient (Chouet 2003) and generate long-period seismic signals. The upward movement of the fluid-particle mixture above the fragmentation surface can trigger vibrations in the conduit structures (Ripepe et al. 2001). Explosions near or within the crater are the sources of seismic signals, as well. The fracturing of lava dome rock may generate seismic signals also (Tuffen et al. 2008).

Waveforms and Spectra

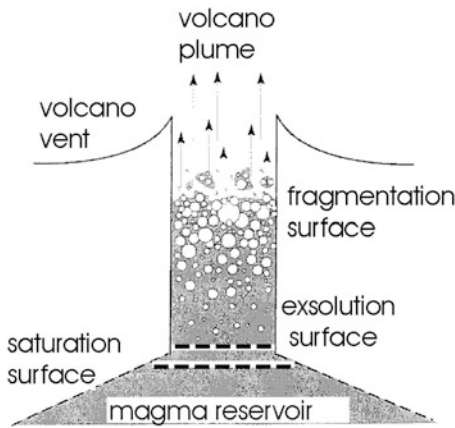
The waveforms of this group of volcanic seismic signals can vary widely depending on the nature of the eruptive process.

Figure 6 shows three typical broadband seismic waveforms associated with *volcanic explosions*. They were recorded at Volcán de Colima, Mexico, and their association with eruptive events was identified with simultaneous video images. All these waveforms are characterized by the existence of large-amplitude, high-frequency signal of the main phase. The main phase may be



Volcano Seismology: An Introduction, Fig. 4 A model for decomposing the deviatoric part of a general moment tensor (**a**) and the focal mechanism solutions of volcano-tectonic earthquakes from the earthquake swarm preceding the 2000 eruption of Miyake-jima volcano (**b**). In (**a**): DC, the double couple source; CLVD, the compensated linear vector dipole. Numbers to the left of the diagrams give the principal moments for each mechanism, in units of 10^{17} N-m. In (**b**): Equal area projection of *P*-wave

first motions is projected on the upper hemisphere. Gray and open area show extensional and compressional quadrant, respectively. Mechanisms of strike-slip type with northeast-southwest stress axis (*P*, most compressive stress; *T*, least compressive stress) are predominant, although various focal mechanism types are recognized. (Figure is compiled using the approach of Julian et al. 1998; Uhira et al. 2005)



Volcano Seismology: An Introduction,
Fig. 5 Schematic illustration of processes that occur in volcanic conduits. (Based on Cashman et al. 2000)

preceded by low-frequency signal *LF* (Type 1, EXP, see Fig. 1), high-frequency signal *HF* (Type 2, HYB), or by both high-frequency and low-frequency signals (Type 3). The spectral peak frequencies of the seismic signals of explosions are recorded between 1 and 2 Hz (Fig. 7).

The *microearthquakes* (ME) (pulgas, drumbeats) are characterized by the waveforms similar to those of explosive earthquakes but of smaller amplitudes and shorter durations. They occur usually in swarms, as shown in Fig. 8a, and may be recorded at different stages of volcanic activity. Two types of microearthquake waveforms were discriminated during the activity of Mount St. Helens, Stromboli, Volcán de Colima, and other volcanoes and were named as long-period and hybrid (Moran et al. 2008; Martini et al. 2007; Arámbula-Mendoza et al. 2011). As seen in Figs. 8 and 9, the waveforms of ME events are similar to the waveforms of explosive earthquakes of Types 1 (EXP) and 2 (HYB), respectively, having low-frequency or high-frequency preliminary phases. The spectra of microearthquakes are characterized by higher frequencies, between 2 and 7 Hz. The volcanic events that generated these two types of waveforms were identified by Zobin et al. (2010) as micro-explosions.

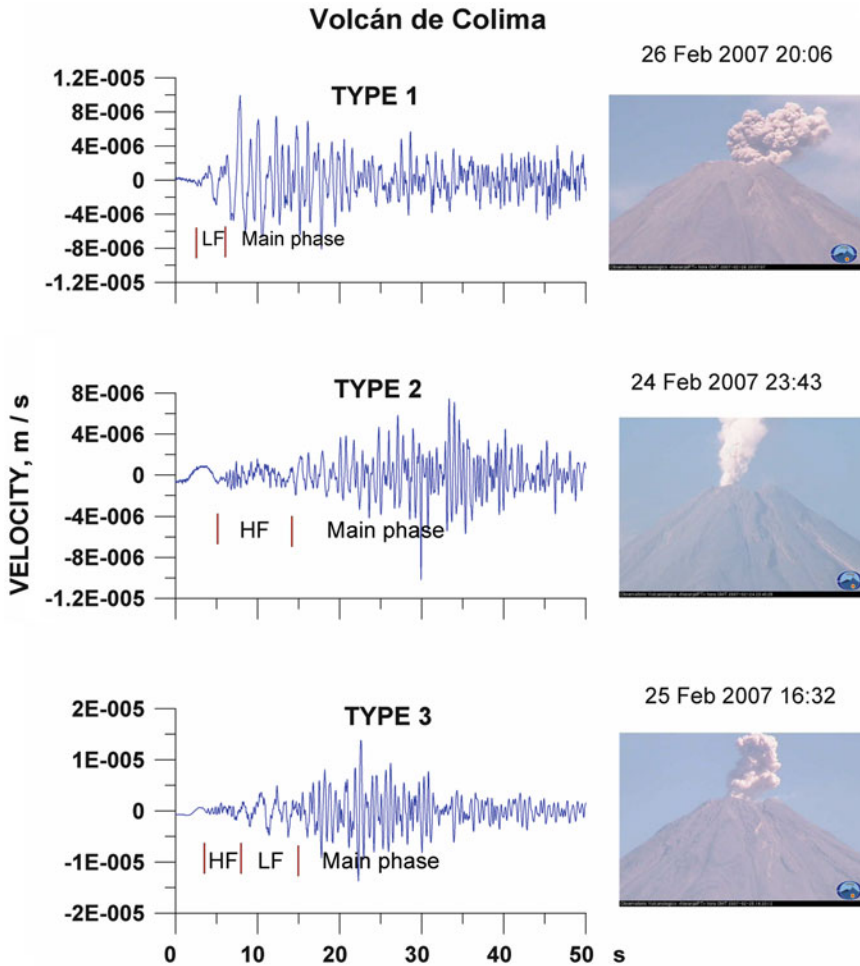
The next group of seismic signals associated with the movement of magma within the conduit represents the signals recorded with an emergent

onset of *P* waves, no distinct *S* waves, and a harmonic signature in the range of periods between 0.2 and 2 s (0.5 and 5 Hz). These events are defined by Chouet (2019) as *long-period* (LP). They are characterized by simple decaying harmonic oscillations except for a brief time interval at the event onset. Seismic signals in the range of periods between 2 and 100 s are defined as *very-long-period* (VLP) (Chouet 2019).

The long-period seismic signals, demonstrated in Fig. 10, were recorded during different stages of eruptive activity. The first of them (Fig. 10a) was observed simultaneously with a swarm of volcano-tectonic earthquakes preceding the 2000 flank eruption at Usu volcano, Hokkaido. It has an emergent high-frequency onset; the body waves are not identified. The amplitude of the signal slowly increases and then gradually decreases. The second signal is the so-called *tornillo* (screw) (Fig. 10c); it was recorded before the 2 January explosive eruption at Galeras volcano, Colombia. Signals of this type show emergent onsets but occasionally may have slight impulsive onsets. They are characterized by slowly decaying coda waves. These events are characterized by a peak spectrum between 0.8 and 2 Hz (Fig. 10b, d).

VLP signals at volcanoes are characterized by impulsive signatures; their common characteristic is a period larger than 2 s (Fig. 11). The VLP signals typically may be discerned only after filtering of the raw signal.

Volcanic tremor represents a continuous record of monotonic, harmonic, or non-harmonic vibrations that may continue for minutes to months. The spectral characteristics of the monotonic and non-monotonic tremors, obtained by Kawakatsu et al. (1992) at Sakurajima volcano, Japan, with a broadband seismometer during the 1991 explosive activity of the volcano, are shown in Fig. 12. It is seen that the short-period spectrum for monotonic tremor is characterized by peaks within a narrow frequency range at about 2.5 Hz and its long-period spectrum is characterized by the absence of any predominant peaks. For non-monotonic tremor, the opposite situation is observed: no unique peaks for the short-period spectrum but a narrow frequency range of peaks at about 0.2 Hz for the long-period spectrum.



Volcano Seismology: An Introduction, Fig. 6 Three types of broadband seismic records (velocity, vertical component) associated with magmatic and gas-and-ash explosions at Volcán de Colima during the 2007 activity. Seismic station was situated at a distance of 4 km from the crater.

Video station was situated at a distance of about 15 km S from the crater. LF and HF indicate preliminary low-frequency and high-frequency signals. (Courtesy of Colima Volcano Observatory)

Volcanic tremor may be represented by surface waves or body waves (Zobin 2017).

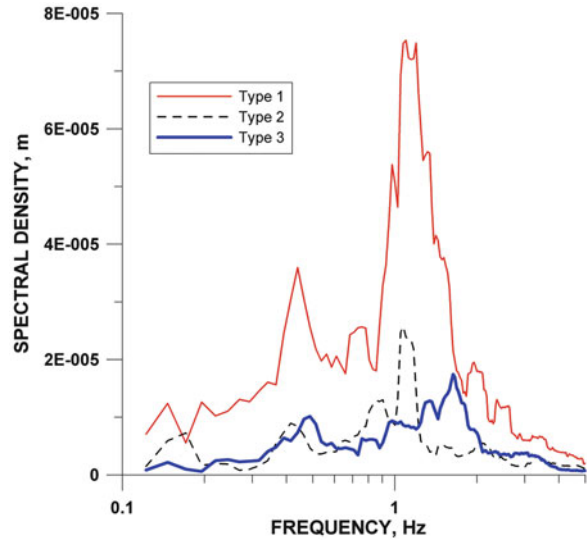
Source Modeling

For modeling of *explosive earthquakes or any long-period seismic vibrations*, the fluid-driven crack model (Chouet 1986), consisting of a rectangular fluid-filled crack contained within an infinite, linear elastic medium, may be applied. This model is described in details elsewhere by Bernard Chouet (2019). Application of this model allows us to calculate the seismic moment

tensors describing volumetric mechanisms representing magma-filled sources of volcanic earthquakes occurring within the volcanic conduit. While the injection of a volcanic fluid or a thermal expansion of a fluid-filled crack occurs, a sudden volume increase, ΔV , is observed, and the fracture that may be modeled as crack, spherical, and cylindrical source acts as a radiating seismic source. The seismic moment tensors of these sources are detailed described in the article of H. Kumagai published in this volume (Kumagai 2019).

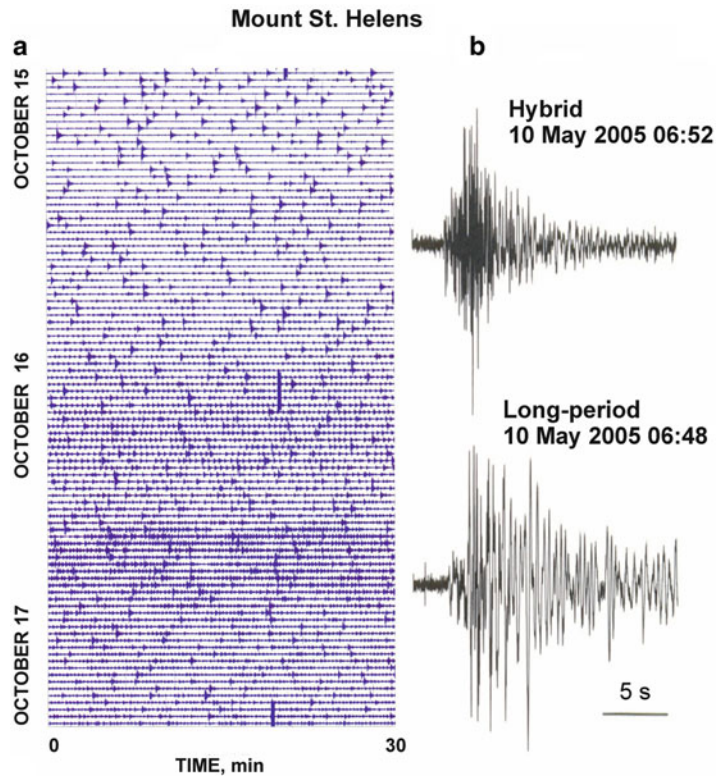
Volcano Seismology: An Introduction,

Fig. 7 Fourier spectra of three types of the seismic signals (vertical component) associated with magmatic and gas-and-ash explosions at Volcán de Colima during the 2007 activity and shown in Fig. 6. (Courtesy of Colima Volcano Observatory)



Volcano Seismology: An Introduction,

Fig. 8 Three-day short-period seismograms of the swarm of microearthquakes (a) and the typical waveforms of the microearthquake signals (b) recorded at a distance of 2.5 km from the crater during the 2004–2005 eruption of Mount St. Helens volcano. (Modified from Moran et al. 2008)

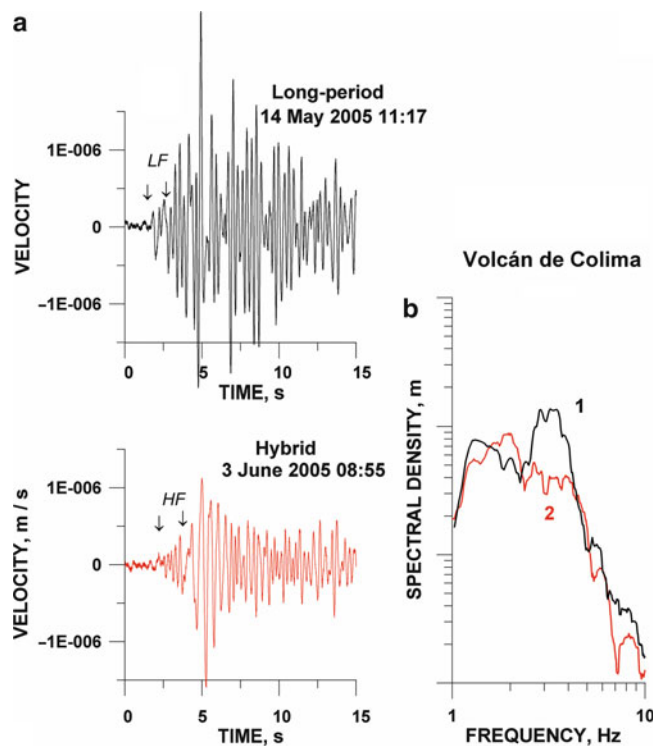


The two-stage conceptual model of explosive process at volcanoes proposed by Zobin et al. (2006, 2009b) describes the volcanic explosion process as consisting of the ascent of fragmented

magma to the surface (stage 1) and a subsequent explosion (stage 2). The comparison of contemporary records of broadband seismic signals and video images of a Vulcanian explosion at Volcán

Volcano Seismology: An Introduction,

Fig. 9 Waveforms (short-period signals, velocity, vertical component) (a) and corresponding Fourier spectra (b) of long-period and hybrid microearthquakes, recorded at Volcán de Colima at a distance of 1.6 km from the crater during the 2005 activity. Arrows show the preliminary phases (*LF* and *HF*) of the seismic records. (Courtesy of Colima Volcano Observatory)

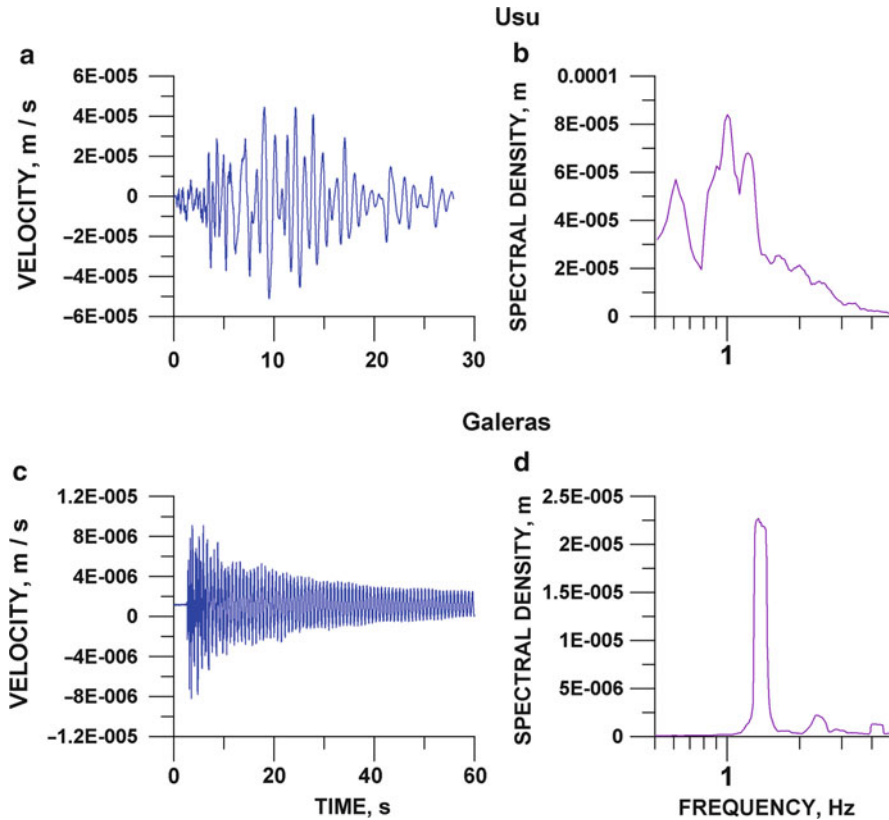


de Colima (Zobin et al. 2008b) shows that the first superficial manifestation of the explosion (visible expulsion of magma material from the crater) appears only about 15 s after the onset of the seismic record, on the coda of the second pulse (Fig. 13). This indicates the ascending process occurring within the volcano conduit before an explosion. Therefore, the seismic record of an explosion (Fig. 13b) may be considered as consisting of two phases, the preliminary pre-explosion phase P_E (between t_1 and t_2 ; it may be LF or HF preliminary phase as shown in Fig. 6) and the co-explosion main phase C_E (between t_2 and t_3). It is supposed that the low-frequency signal of the preliminary phase may be generated during the passage of magma jet within the conduit and the high-frequency signal may be produced by the fracturing of the magma body. Application of this model allows to calculate the counter force of eruption from waveform inversion of the pre-explosion phase and the energy of eruption of Strombolian and Vulcanian explosions from Fourier spectrum of the co-explosion main phase (Zobin et al. 2009a). This model may

be applied also for size estimation of microearthquakes (Zobin et al. 2010).

The relative source process generating the LP and VLP events that were located using waveform inversion was discussed by Waite et al. (2008) for the 2 July 2005 Mount St. Helens event. In the seismic record, the main part of the VLP signal occurred after the initiation of an LP event. The position of the LP and VLP sources is shown in Fig. 14. The LP source is very shallow, at about 2000 m elevation and directly beneath the growing dome. The VLP source is 250 m deeper and 400 m NNW, beneath the SW end of the old 1980s lava dome.

A possible scenario for the LP events at Mount St. Helens involves the repeated pressurization of a subhorizontal crack that lies beneath the southern crater and the new lava dome. Steam generated by the heating of groundwater, in addition to exsolved steam from the magma, fed into the crack, gradually expanding and pressurizing it. Waveform inversion of the VLP events revealed a north-dipping sill plus a near-vertical dyke or pipe. The VLP source sill was roughly in



Volcano Seismology: An Introduction,
Fig. 10 Examples of long-period seismic waveforms (velocity, vertical component) and their Fourier spectra. (a) Seismic signal recorded at Usu volcano on 30 March

2000 (Courtesy of Japan Meteorological Agency) and its Fourier spectrum (b). (c) Seismic signal recorded at Galeras volcano on 1 January 2010 (Courtesy of INGEOMINAS, Colombia) and its Fourier spectrum (d)

the plane of the LP source crack. The main VLP signal is supposed to be generated by deflation-inflation of the sill and near-simultaneous inflation-deflation of the dyke.

Among another possible sources of LP and VLP signals may be mentioned the magmatic-hydrothermal interactions (Chouet and Matoza 2013).

Many models were proposed to explain the seismic signals of *volcanic tremor*. One of them is the conduit-vibration model (Schick et al. 1982), which this model hypothesized that transient flows within volcanic conduits can generate tremor with certain dominant frequencies, depending on the conduit geometry. Vapor and gas escaping through open conduits are considered responsible for turbulent motions in the magma; consequently the magma-filled pipes in

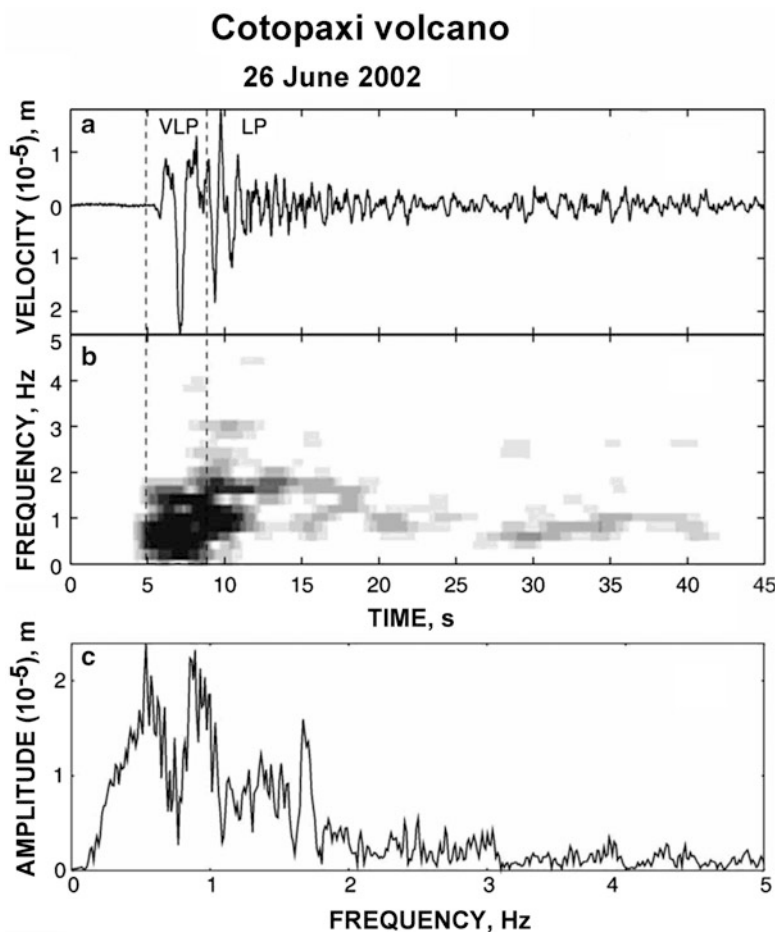
the upper structures of the volcano behave as oscillators. Figure 15a shows a model of conduit oscillators for Etna volcano, Sicily. It consists of a set of vibrators with characteristic frequencies from 0.9 to 4 Hz. These characteristic frequencies are comparable with the peak frequencies that were observed for volcanic tremor at Etna (Fig. 15b).

Seismic Signals Associated with the Surface Manifestations of Volcanic Activity

The motions of subaerial masses of sediments at volcanoes are divided into two groups: the phenomena in which the particles are dispersed in the flowing body and the phenomena in which the

Volcano Seismology: An Introduction,

Fig. 11 Broadband velocity VLP seismic record accompanied by LP signal (a) recorded at Cotopaxi volcano on 26 June 2002 without any reported surface manifestation of volcanic activity, its spectrogram (b), and Fourier spectrum (c). (Modified from Molina et al. 2008)



moving bodies are mostly the agglomerates of soil and rocks (Takahashi 2007). The pyroclastic flows and rockfalls are included in the first group; lahars belong to the second group.

The gravitational collapse (Merapi type) or the explosive disruption of a growing lava dome (Peléean type) and the collapse of an eruption column (Soufrière type) can generate pyroclastic flows and rockfalls (Francis 1998). Pyroclastic flows are distinguished from rockfalls by their larger size, longer runout (≥ 0.5 km), greater production of fine ash, and appreciable buoyant hot ash clouds (Calder et al. 2002). Lahars along the slopes of volcanoes are usually triggered by melting of snow/ice or heavy rainfall, especially after the intensive fall of fresh, still unconsolidated volcanic debris. In some instances, lahars may have been triggered by steam explosions that

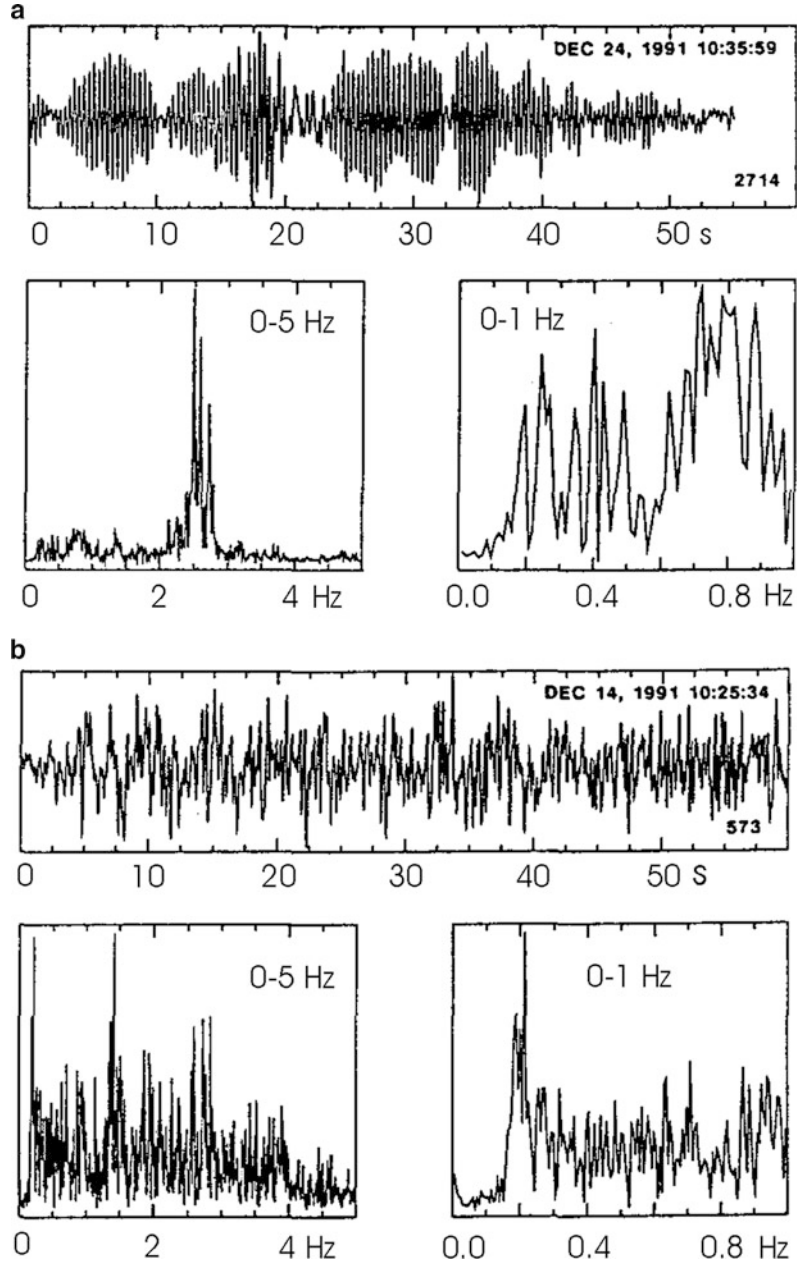
generated strong ground vibrations, which induced failure of partially water-saturated pyroclastic materials (Rodolfo et al. 1996). The movement of pyroclastic flows, rockfalls, and lahars along the volcanic flanks generates seismic signals.

Waveforms and Spectra

Figure 16 shows broadband seismic signals associated with the pyroclastic flows generated by two types of volcanic processes at Volcán de Colima during its 2004–2005 eruption. Their association with eruptive events was identified with simultaneous video images and photo. The signal recorded on 1 October 2004 (Fig. 16a) was associated with a pyroclastic flow generated due a partial collapse of a lava dome during the lava extrusion (Merapi type). The waveform begins

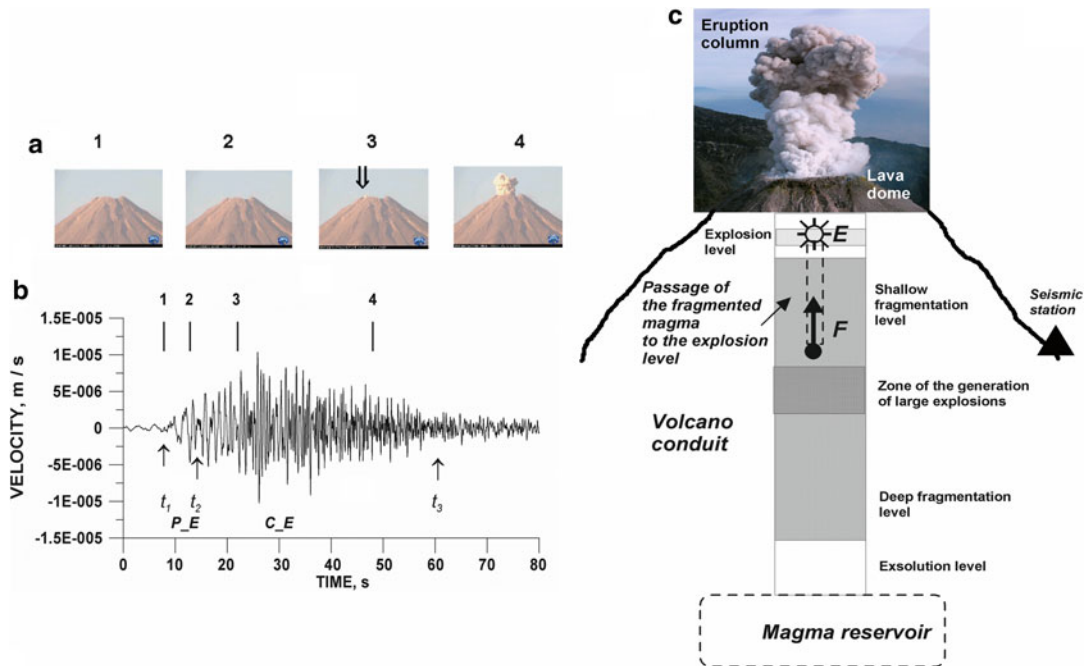
Volcano Seismology: An Introduction,

Fig. 12 Vertical velocity broadband records of monotonic (a) and non-harmonic (b) tremors observed at Sakurajima volcano, Japan. The initial time of the records and peak-to-peak amplitude (in 10^{-8} m/s) of each seismogram are given at the upper-right and lower-right corners. Vertical scales of spectra are arbitrary. (From Kawakatsu et al. 1992)



with a low-amplitude signal which gradually increases in amplitude and then more slowly decreases. The second signal recorded on 13 March 2005 (Fig. 16b) was associated with the pyroclastic flow that accompanied large Vulcanian explosion and was produced by the collapse of an eruption column formed by the explosion (Soufrière type). Its waveform is more

complex. The seismic record begins with the large-amplitude signal produced by explosion (EXPL in Fig. 16b); then the collapse of volcanic material from an eruption column generates pyroclastic surge (PF). The initial seismic signal of an explosion has significantly larger amplitudes than the signal generated by a pyroclastic flow that make difficult to see exactly the seismic signal

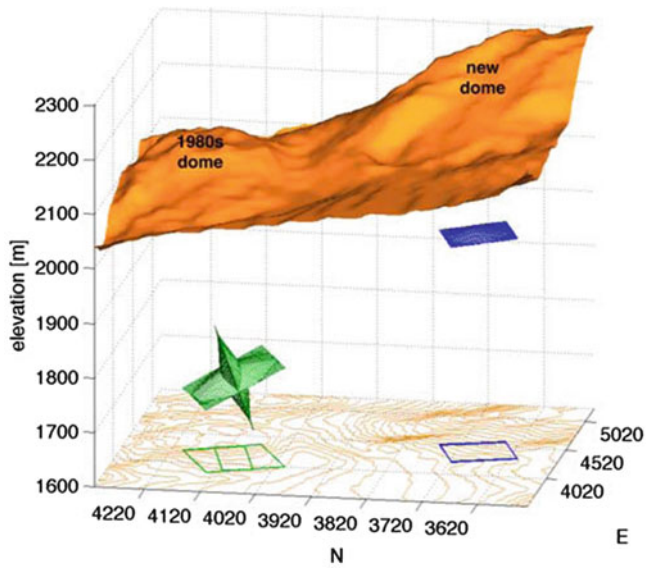


Volcano Seismology: An Introduction,
Fig. 13 Comparison of broadband seismic record (EZ5 station) of an explosion at Volcán de Colima (21 February 2007) with the video images of the same event. 1–4 corresponds to the timing of images and seismic record; t_1 , t_2 , and t_3 are the times of the onset and the end of pre-explosion (P_E) pulse and the end of co-explosion (C_E)

pulse, respectively; the arrow shows the start of ejection of explosive material from the crater. The digital infrared video camera was installed at a distance of 15 km from the crater and records a frame/second. The timing at seismic station and for the video camera was synchronized by the world atomic time survey. (From Zobin et al. 2009b)

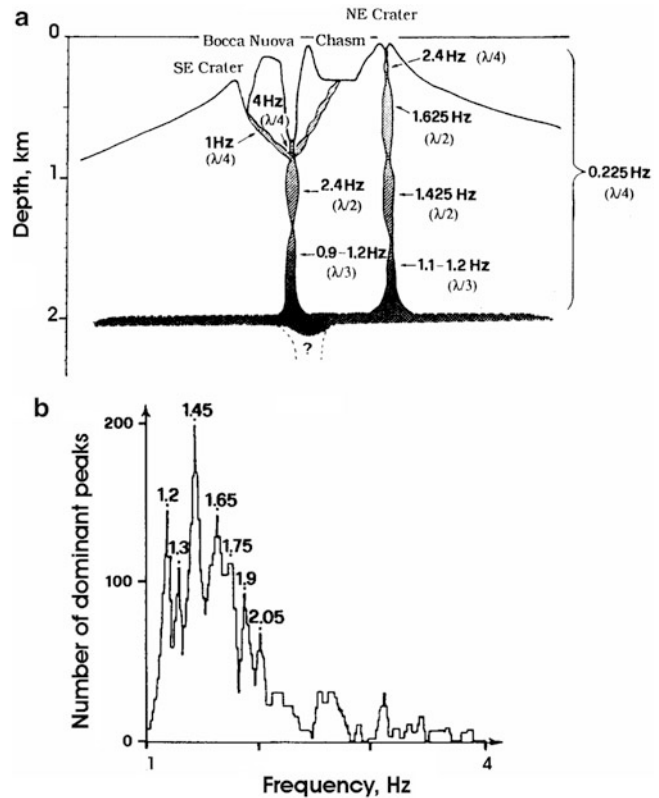
Volcano Seismology: An Introduction,

Fig. 14 Relative position of the sources that generated the 2 July 2005 LP and VLP seismic signals during lava dome growth in the crater of Mount St. Helens. The east-northeast, three-dimensional, perspective view of the LP crack (blue) and VLP sill-dyke (green) sources are shown. Topography is shown at the bottom as a surface and is contoured at 20 m. Distances are in m. (From Waite et al. 2008)



Volcano Seismology: An Introduction, Fig. 15 (a)

Schematic model of the main magma conduits at Etna. Bocca Nuova and Chasm are active craters of Etna. (b) Histogram of the first three dominant peaks of tremor recorded from January 1973 to January 1985 at Etna. (Compiled from Schick et al. 1982; Cosentino et al. 1984)



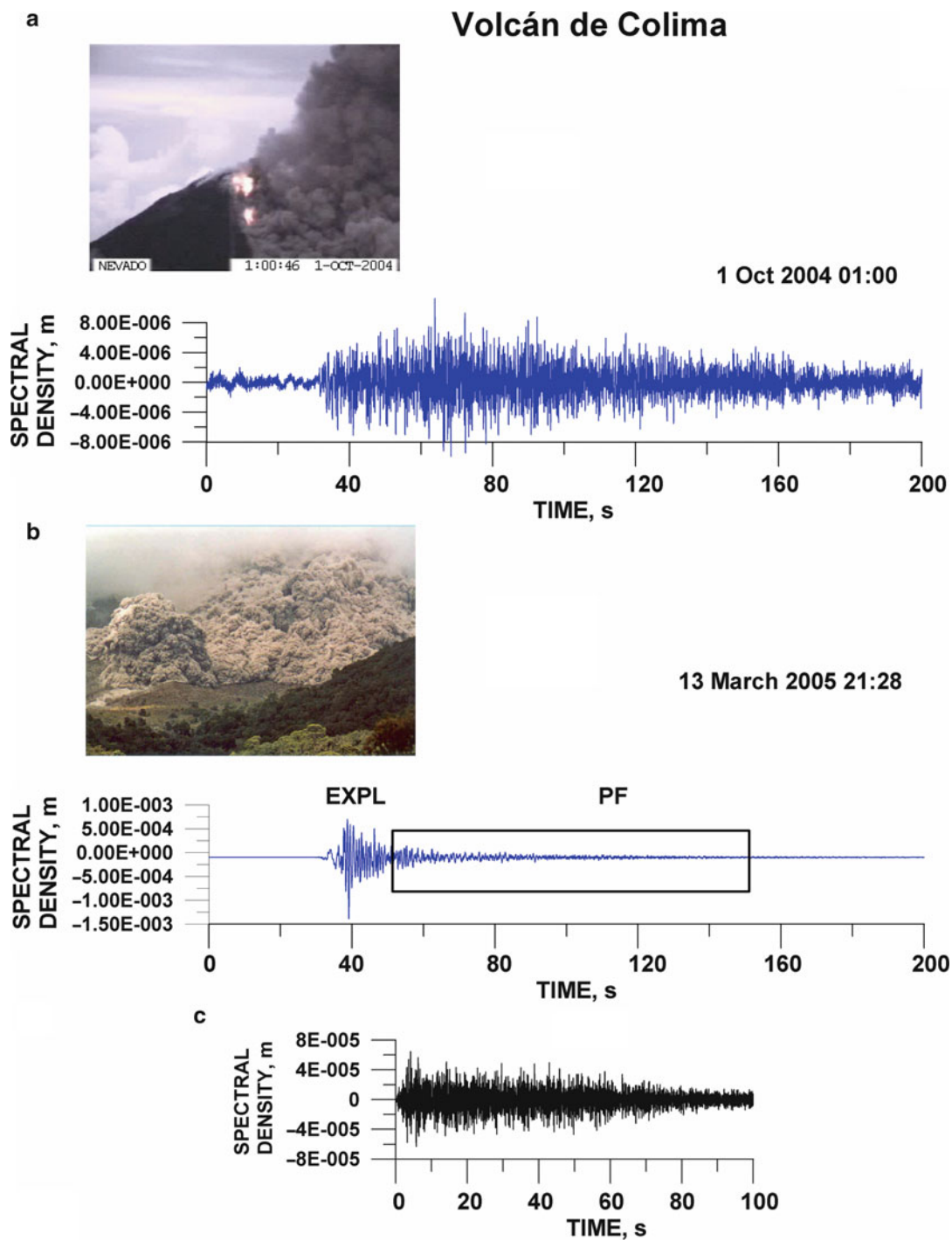
associated with a pyroclastic flow. To reach it, a 100-s portion of the seismic record of Fig. 16b, following the explosive onset, was selected and shown separately in Fig. 19c. In this presentation, the seismic signals associated with two types of pyroclastic flows are looking similar enough. Fourier spectra of these seismic signals are shown in Fig. 17. The spectra of both types of pyroclastic flows (PF_collapse and PF_explosion) are similar with close peak frequencies between 2 and 5 Hz.

As noted by Calder et al. (2002), pyroclastic flows are distinguished from rockfalls by their larger size, longer runout (≥ 0.5 km), greater production of fine ash, and appreciable buoyant hot ash clouds. The seismic signal associated with a rockfall, recorded during the 1997 eruption activity at Soufrière Hills volcano, Montserrat, is shown in Fig. 18. As is seen, its waveform is similar to those produced by pyroclastic flows (Fig. 16). The peak frequency of the rockfall signal spectrum is equal to 6 Hz, higher than for pyroclastic flows signals.

The 1-h seismic records produced by the 2005 lahars passing along the ravines at Volcán de Colima are shown in Fig. 19a. They represent a sequence of impulses that gradually increase in time, reaching their maximum amplitude, and then gradually, but more slowly, decrease. Figure 19b shows Fourier spectra calculated for these broadband seismic records of lahars. Their peak frequencies are between 6 and 8 Hz.

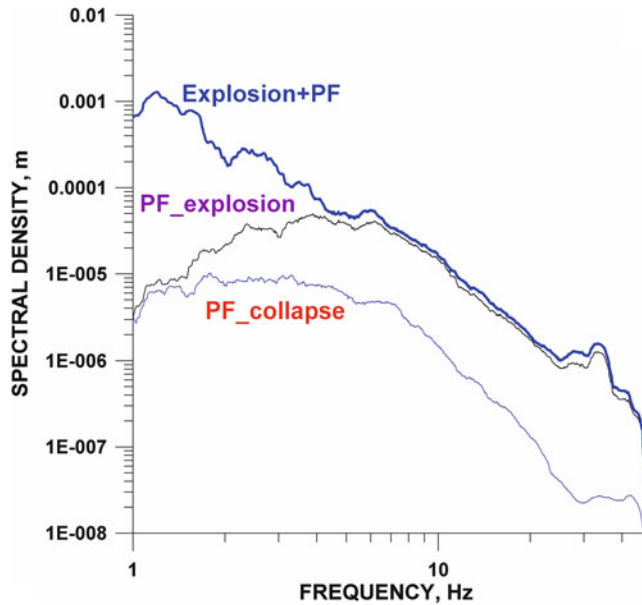
Source Modeling

Figure 20 shows the mechanical models proposed for modeling of pyroclastic flows and lahars. The propagation of pyroclastic flow can be represented as a three-stage process (Takahashi and Tsujimoto 2000): the collapse of lava dome (or the rocks from the eruptive column) induces a granular flow, which later changes to a fluidized flow (Fig. 20a). The observations of generation of pyroclastic flows at Unzen volcano, Japan (Yamasato 1997; Uhira et al. 1994) illustrated these three stages in the formation of a pyroclastic



Volcano Seismology: An Introduction, Fig. 16 (a) Video image of the pyroclastic flow produced by the partial collapse of lava dome at Volcán de Colima (1 October 2004) and broadband seismic record (vertical component) of this pyroclastic flows. (b) *Photo of the pyroclastic flow*

associated with the collapse of volcanic material from the volcanic column during the 13 March 2005 explosion at Volcán de Colima and broadband seismic record (vertical component) of the explosion (EXPL) and the following pyroclastic flow (PF). Both seismic signals were recorded



Volcano Seismology: An Introduction, Fig. 17 Fourier spectra of broadband seismic signals associated with pyroclastic flows recorded during the effusive-explosive activity at Volcán de Colima and shown in Fig. 16. The spectra shown are of the signal associated with the pyroclastic flows produced by the partial collapse of lava dome (PF_collapse), the signal

including the seismic record of the explosion and the following pyroclastic flow produced by the collapse of volcanic material from the volcanic column during the explosion (Explosion + PF), and the part of this signal associated with the pyroclastic flow only (PF_explosion). (Courtesy of Colima Volcano Observatory)

flow, recorded by video and seismic instruments (Fig. 21a): the lava dome collapse (stage 1), the downfall of lava blocks onto the slope (stage 2), and the generation of pyroclastic flow (stage 3). This process may be described by a function representing the single-force system that is shown in Fig. 21b (Uhira et al. 1994).

Lahars, in their turn, may be described as hybrid-type debris flows between turbulent muddy and stony debris flows (Takahashi and Satofuka 2002). The hybrid flow in the inertial range consists of two layers (see Fig. 20b): the upper layer of depth h_1 is the particle suspension layer, and the lower layer of depth h_2 is where particles move with frequent collision or enduring contact with each other. The

dominance of stony or turbulent muddy type of flow depends on the relative thickness of these two layers h_1/h_2 . The movement of the lower stony layer along the volcanic slope ravine (angle of inclination θ in Fig. 20b) produces the main part of the seismic signal of lahar.

Relationship of Volcanic Earthquakes with Eruption Process

The Swarms of Volcano-Tectonic Earthquakes as Forerunners of Volcanic Eruption

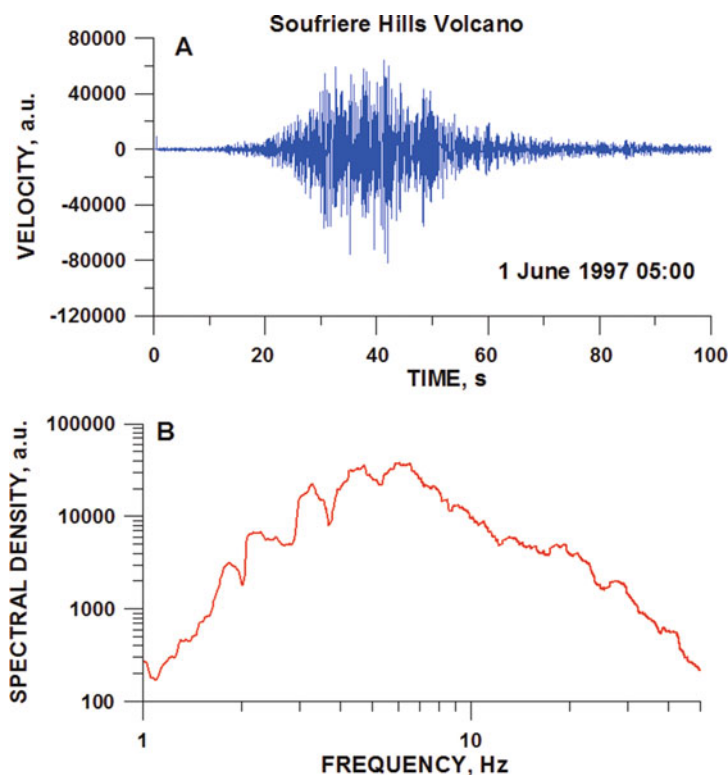
Swarms of volcano-tectonic earthquakes may occur before the eruption, during the eruption,

Volcano Seismology: An Introduction, Fig. 16 (continued) at a distance of 4 km from the crater. The bold-line rectangle on the seismogram of figure (b) shows the selected sample (see text) presented in figure (c). To

remove the low-frequency noise of explosion, this signal was high-pass filtered at 2.5 Hz. (Courtesy of Colima Volcano Observatory. Photo of the explosion in B was taken by Sergio Velasco)

Volcano Seismology: An Introduction,

Fig. 18 Broadband seismic record (velocity, vertical component) associated with rockfall that was observed at Soufriere Hills volcano on 1 June 1997 during the 1997 effusive-explosive episodes and recorded at a distance of 4 km from the crater (a) and its Fourier spectrum (b). The event was recorded at a distance of 2.5 km from the dome. (Courtesy of British Geological Survey, Edinburgh, UK)

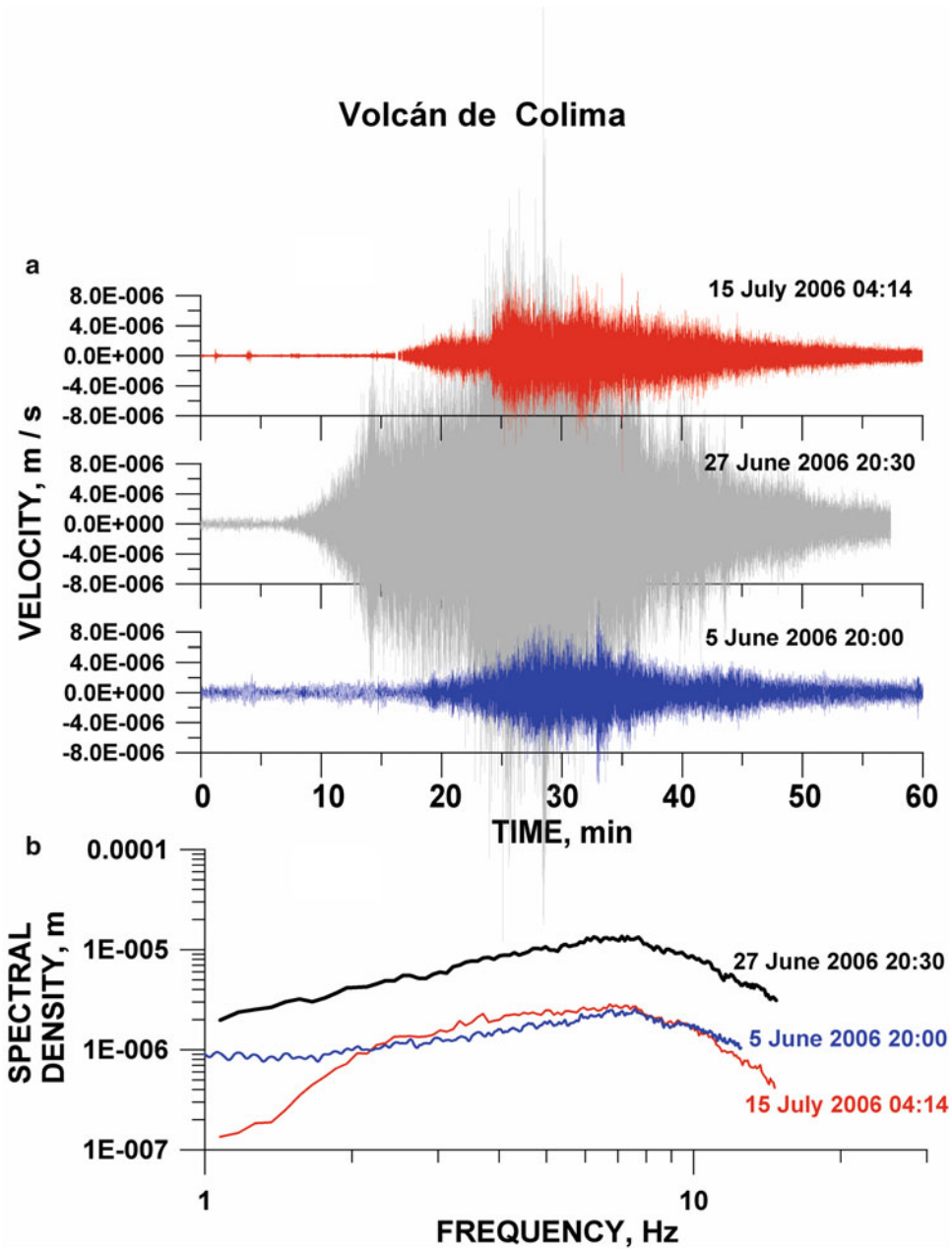


and after the eruption of volcano; such earthquakes also are typically associated with dike intrusions that do not culminate in eruption. The development of the swarms of volcano-tectonic earthquakes depends on the type of eruption or magma intrusion (summit, flank, or fissure) and on the type of magma feeding the eruption (basaltic, andesitic, or dacitic). The duration of seismic swarms prior to an eruption of the majority of andesitic and dacitic volcanoes ranges between 10 and 1000 h. For majority of basaltic volcanoes, the durations of seismic activity prior to eruption are shorter (from 1 to 100 h). Volcanic eruptions may occur during the different stages of seismic sequence. For basaltic volcanoes, the majority of eruptions occur after the end of the seismic sequence, while for andesitic volcanoes, the majority of eruptions occur during the maximum frequency of volcano-tectonic earthquakes (Zobin 2017).

Very commonly the occurrence of a swarm of volcano-tectonic earthquakes near a volcano may indicate a forthcoming eruption of the volcano.

The 2000 lateral eruption of Usu volcano, Hokkaido, was the first case in Japan when a formal alert was issued by the Japan Meteorological Agency (JMA) before the eruption (Okada 2002; JMA 2003). The 2000 eruption began on the northwestern flank of the volcano (Fig. 22) on 31 March, following about 4 days of intensive seismic activity and a high-rate deformation of the volcano surroundings. About 1200 earthquakes were located by Japan Meteorological Agency (JMA) from 28 March to 1 April, 339 of them with magnitudes ranging from 2.5 to 4.6.

During the first 40 h of activity, the frequency of events was rather low, not exceeding ten located events per hour (Fig. 23). The maximum magnitudes of earthquakes were 3.2–3.4. The number of earthquakes and their magnitudes increased sharply at the end of 29 March and reached their maximum between 9 and 12 h of 30 March. Then seismic activity gradually decreased and practically ceased on 2 April. Three largest earthquakes of the swarm occurred



Volcano Seismology: An Introduction,
Fig. 19 Broadband seismic records (velocity, vertical component) of lahars propagating along the ravines on the

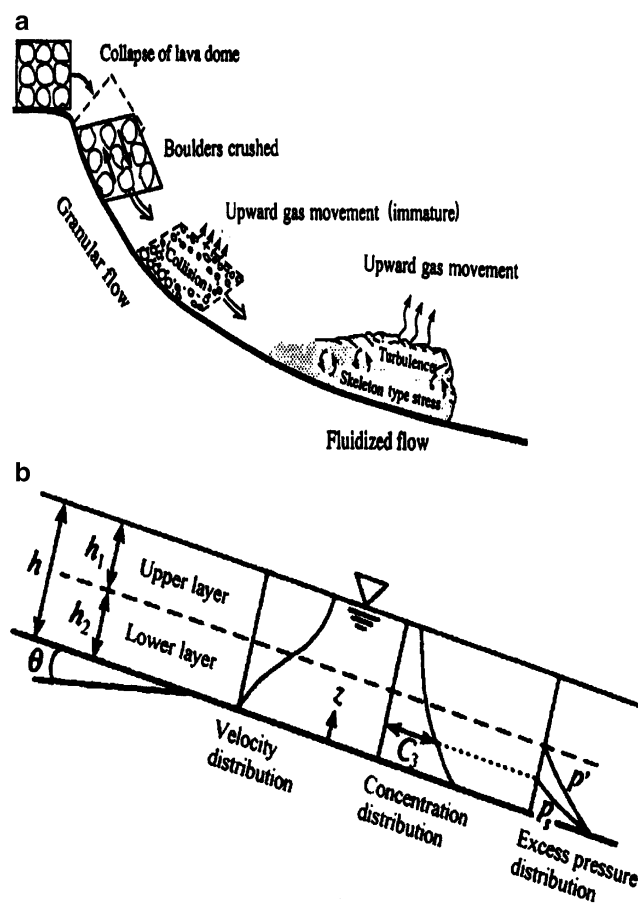
flanks of Volcán de Colima (a) and their Fourier spectra (b). The seismic signals were recorded at a distance of 4 km from the crater. (Courtesy of Colima Volcano Observatory)

during the maximum seismic energy release (M 4.4, 30 March, 09:12 and 17:13) and at the final stage of the swarm, 14 h after the beginning of the eruption (M 4.6; M_w 4.9, 1 April, 03:12) (Zobin et al. 2005a).

The earthquake epicenters occupied an area of about 30 km^2 , covering the summit zone and southern and southwestern flanks of the volcano (Fig. 22). The foci were recorded within the depth range between 0 and 13 km but clustered mainly

Volcano Seismology: An Introduction,

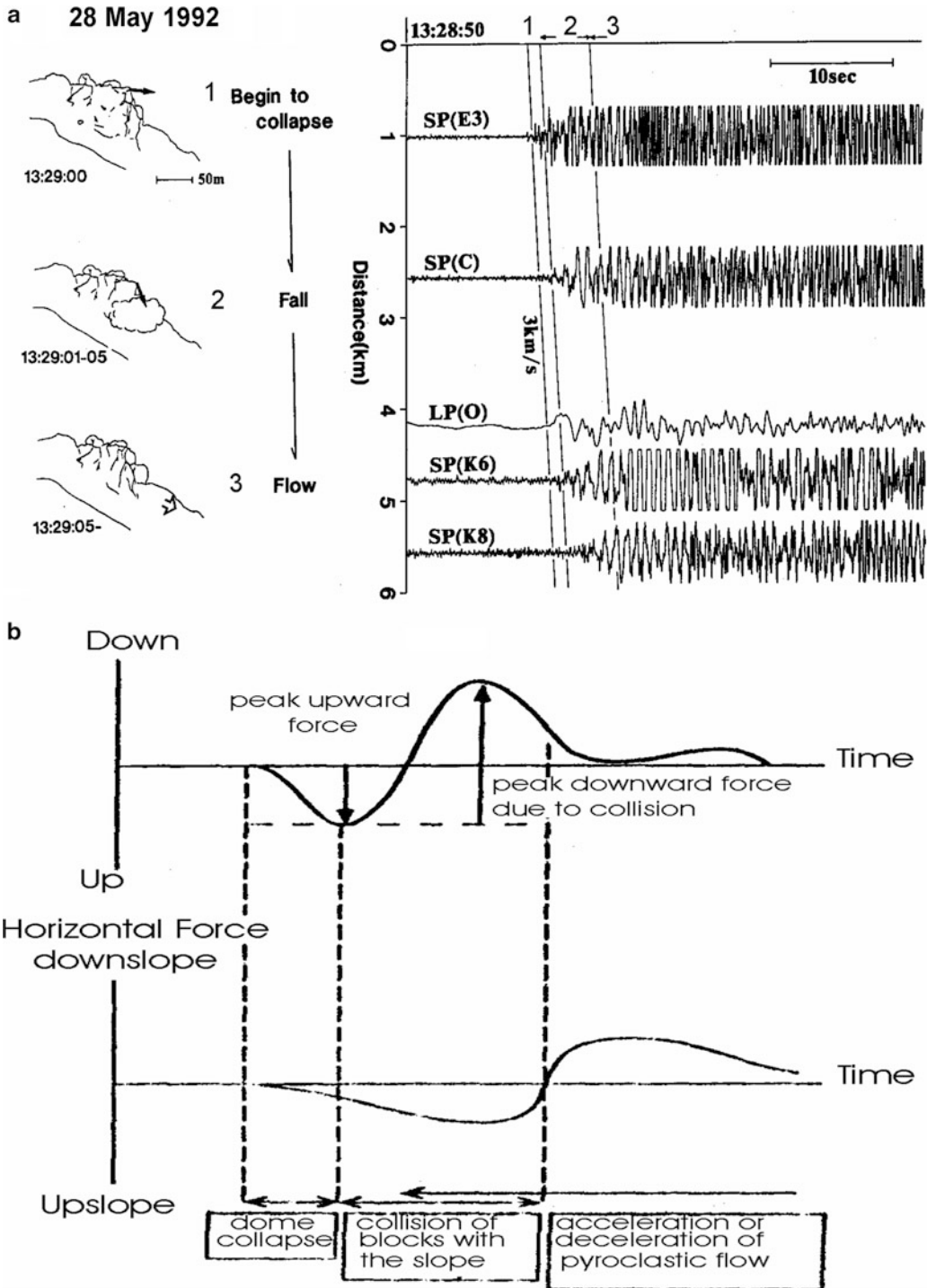
Fig. 20 Mechanical models of pyroclastic flows (a) and lahars (b). Details are discussed in text. After Takahashi and Tsujimoto (2000) and Takahashi and Satofuka (2002)



from 4 to 7 km. The earthquake epicenters stopped at a distance of about 1–3 km from the sites of forthcoming eruption.

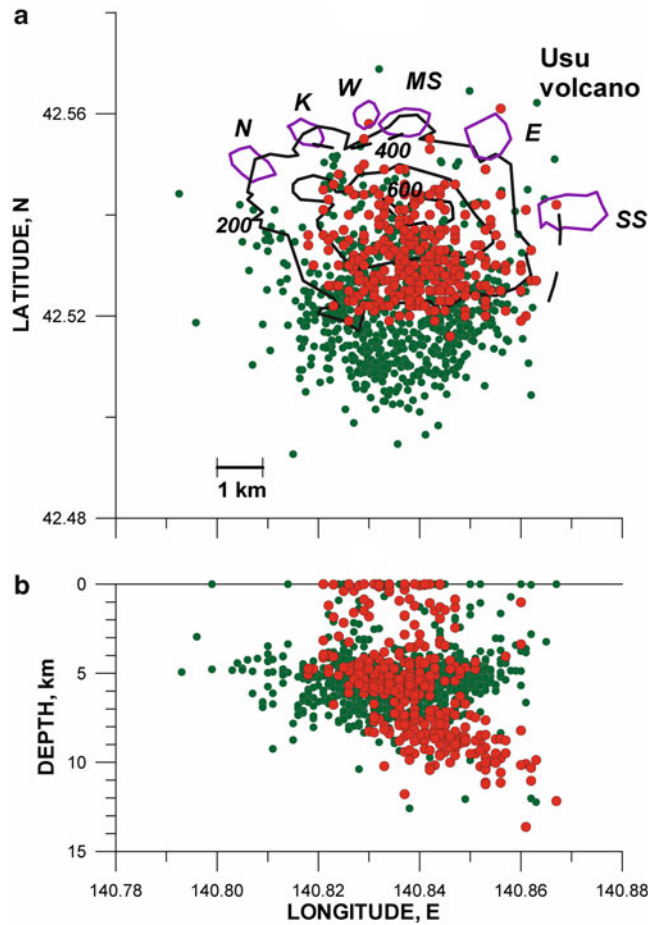
The process of forecasting of this eruption, basing on the development of the swarm of volcano-tectonic earthquakes (Okada 2002), is illustrated in Fig. 23. When the number of volcano-tectonic earthquakes, recorded at the volcano, increased clearly during the morning of 28 March, the Muroran Local Meteorological Observatory issued the first report about this situation, the Volcanic Advisory about the warning attention. Emergency meetings of the Coordinated Committee for Prediction of Volcanic Eruption of JMA were held on 28 and 29 March to evaluate all data and information. GPS observations indicated that the volcano had inflated since March 29. The Committee concluded that the eruption of Usu volcano

should be considered “imminent” within a few days. Based on this conclusion, a “Volcanic Alert” was issued at 11:10 (Japan Standard Time) 29 March by JMA. This alert signified that volcanic activity reached extremely high levels and urgent measures to prevent and mitigate danger to human lives are required. Based on Volcanic Alert of 29 March, local governments established a headquarters for volcanic disaster prevention and issued evacuation orders. Fifteen thousand residents and tourists were evacuated from the area of volcanic danger by local governments without any confusion. After 30 March, cracks and faults were also found in the northwestern foot and summit, indicating that crustal deformation had increased (JMA 2003). The last alert was issued on 31 March, a few hours before the eruption.



Volcano Seismology: An Introduction, Fig. 21 (a) Illustrations of the dome collapse sequence at Unzen volcano, Japan (left), and the corresponding vertical components of seismic signals (right). A sketch of the dome from the seismic station, situated about 3 km to S of the dome, is

shown based on a video record taken by a portable 8 mm video recorder. SP indicates the short-period records; LP, the long-period records. The records are arranged according to the horizontal distance from the source. E3, C, O, K6, and K8 are seismic stations. (b) Schematic



Volcano Seismology: An Introduction, Fig. 22 The epicenters (a) and hypocenters (b) of earthquakes (JMA catalog) that occurred at Usu volcano from 28 March to 1 April 2000. The events of 28–29 March are shown by red symbols; the events of 30–31 March are shown by green symbols. The 200, 400, and 600 m contours of the Usu volcano are shown by lines. The centers of flank eruptions of Usu volcano are shown by the ovals with letters. *N* and

K are the 2000 centers; *MS* and *SS* are the centers of eruptions during 1910 and 1944, respectively; the centers *W* and *E* were not dated. The broken lines show the lateral extent of volcanic activity during the 1910 and 1943–1944 flank eruptions. The positions of flank eruptive centers and the lines of lateral extent of volcanic activity are shown according to Yokoyama and Seino (2000). (Data from the catalog of Japan Meteorological Agency)

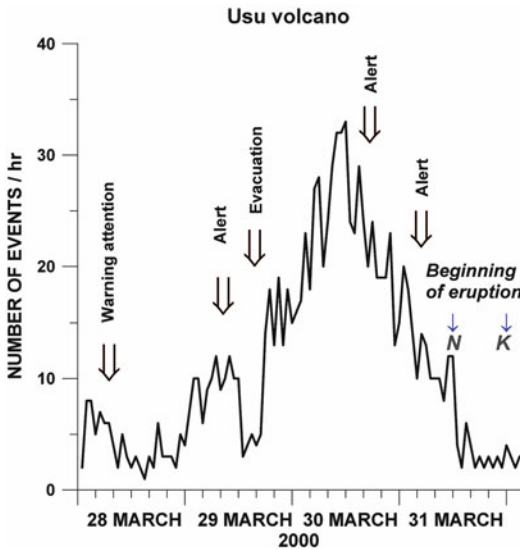
Among other successful predictions of volcanic eruptions may be mentioned the 1975 fissure Tolbachik eruption in Kamchatka (Tokarev 1978); the 1991 explosive eruption at Pinatubo, the Philippines (Harlow et al. 1996); and the 1998 summit eruption at Volcán de Colima, México (Reyes-Dávila and De la Cruz-Reyna 2002).

Seismic Events That Accompany Volcanic Eruption as the Source of Information About Dynamics of Eruption

Numerous seismic events of different types occur during volcanic eruption, and they can provide diagnostic information about eruptive processes.

Volcano Seismology: An Introduction, Fig. 21 (continued) source time functions of the vertical and horizontal components of a single-force exciting pyroclastic flow

according to the dome collapse sequence shown in figure (a). (Compiled from Yamasato 1997; Uhira et al. 1994)



Volcano Seismology: An Introduction, Fig. 23 Temporal developments of the 2000 earthquake swarm at Usu and a sequence of alarms before the 2000 Usu eruption. Data are taken from the catalog of Japan Meteorological Agency and are shown for Japan Standard Time (JST = GMT + 9 h). A sequence of alerts (double arrows) is shown according to Okada (2002). N and K with thin arrows indicate the time of eruptions from two new centers, Nishi-yama and Kompira-yama

Volcanic Tremor

Important information about eruptive processes, mainly for basaltic volcanoes, may be obtained from analysis of the seismic signals of volcanic tremor, for a short episode and for the total eruption. Figure 24 shows the relationships between the intensity of volcanic tremor and eruptive events. A comparison between volcanic tremor amplitude and video camera frames of the 29 May 2000 lava fountain episode at Etna volcano (Alparone et al. 2003) is illustrated in Fig. 24a. An increase and decrease in the amplitudes of the seismic signals are seen depending on the height of the lava fountain. The maximum tremor amplitude was reached at 19:55, after which the height of the fountain halved in no more than 2 min and the tremor markedly decreased in 5 min.

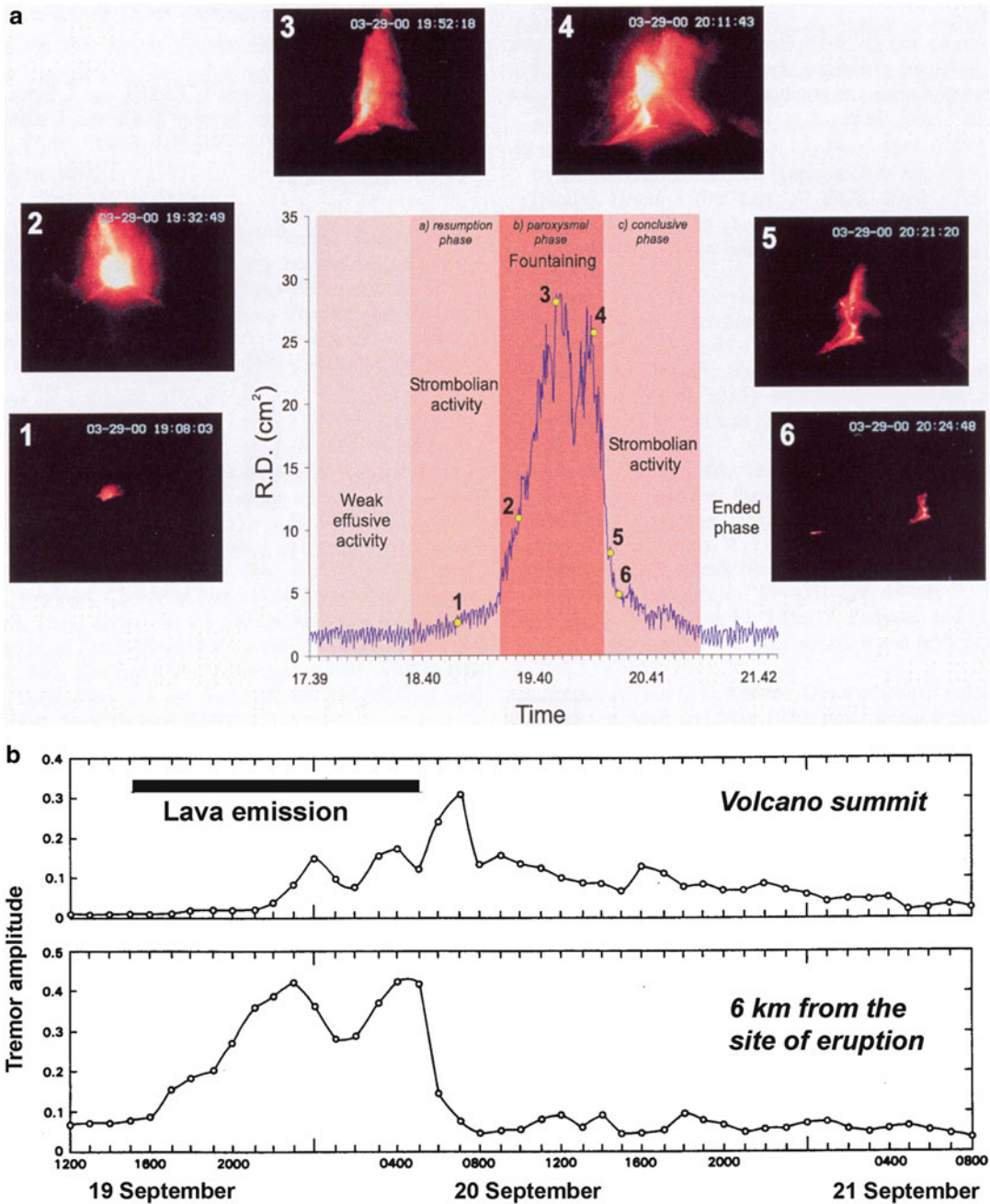
During September 1984, the ongoing long-lived Puu Oo eruption (1983–2018) on the east rift zone of Kilauea volcano was marked by a brief

episode of very high lava fountaining (maximum height 467 m). Figure 24b demonstrates that the tremor began with the start of the high fountaining, gradually reached its maximum value over 7 h, and stayed at this level until the vigorous fountaining ceased. A rapid decrease of tremor then accompanied the abrupt decline of lava emission and fountaining activity (Koyanagi et al. 1987).

Pyroclastic Flows and Rockfalls

A good example of relationship between the effusive activity and variation in the number of rockfalls was observed during the 2004 andesitic lava extrusion at Volcán de Colima (Fig. 25). An intensive swarm of seismic events produced by rockfalls and pyroclastic flows indicated the overflow of lava on 30 September with the formation of two-block lava flows (Zobin et al. 2008a). The emission of lava reached its maximum rate of discharge on 6 October; it then continued with a decreasing rate until practically terminating in November. The variation in the number of seismic signals of rockfalls, recorded at a distance of 1.6 km from the crater, is correlated with lava extrusion dynamics; their peaks coincide. With termination of lava emission, the number of rockfalls sharply decreased.

The amplitudes and durations of the seismic signals produced by rockfalls and pyroclastic flows are good correlated with the volume of deposits of rockfalls and pyroclastic flows that may be useful for monitoring of effusive eruption and approximate estimation of the deposits volume. Thus, a mean volume of $2.0 \times 10^5 \text{ m}^3$ was obtained for the pyroclastic flows, occurring on Volcán de Colima, when the seismic signal duration was greater than 250 s, and a volume of $1.0 \times 10^3 \text{ m}^3$ as a mean value for the events with the seismic signal duration between 100 and 250 s (Zobin et al. 2005b). The data, presented by Uhira et al. (1994) for pyroclastic flows of Unzen, gave the following relationship between the volume of the blocks fallen from the dome, V , and the maximum amplitude of long-period seismic records, A_{max} , produced by pyroclastic flows:

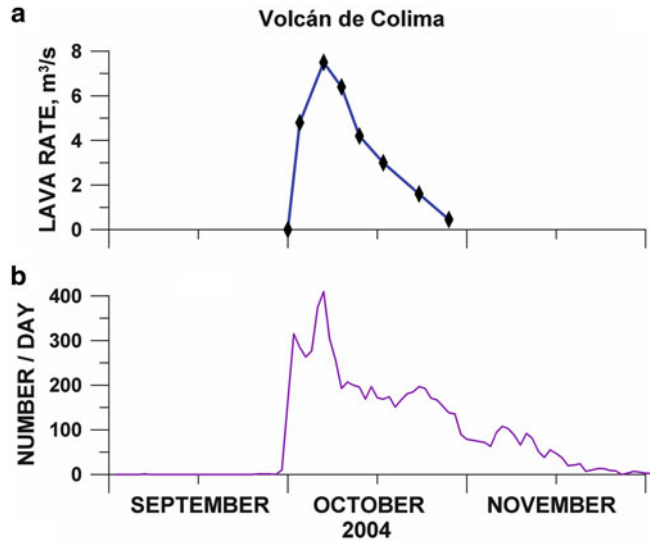


Volcano Seismology: An Introduction, Fig. 24 (a) Comparison between volcanic tremor amplitude and video camera frames of the 29 March lava fountain episode at Etna volcano. (b) Variation in the tremor peak-to-peak amplitudes (in micrometers) during and after an episode of high lava fountaining in September 1984 during the Puu

Oo east rift eruption at Kilauea volcano, Hawaii, as recorded at the volcanic summit (above) and at a distance of 6 km west from the site of fountaining (below). The duration of vigorous fountaining and lava emission is shown. (Compiled from Alparone et al. 2003; Koyanagi et al. 1987)

Volcano Seismology: An Introduction,

Fig. 25 Variations in the lava emission rate (a) and the number of rockfalls (b) recorded at short-period station EZV4 during the 2004 extrusion at Volcán de Colima. (Courtesy of Colima Volcano Observatory)



$$V (10^5 \text{ m}^3) = 0.22 A_{\text{max}} (\text{micron}) + 0.06$$

For both volcanoes, the seismic signals were recorded at a distance of 4 km from the crater.

Volcanic Explosions

Seismic signals associated with volcanic explosions may help to quantify the size of explosions and to understand their nature. The goal of quantifying the size of explosive volcanic eruptions has produced several magnitude and intensity scales, including the eight-grade Volcanic Explosivity Index (Newhall and Self 1982). These scales are based primarily on the amount of volcanic material produced by a sequence of explosions during the eruption and cannot give the size estimation for each of explosions representing this eruption. The initiation of broadband seismic observations at volcanoes has allowed the quantification of an individual explosion of the eruption using physical parameters such as the counterforce of the eruption, the moment tensor components, and the energy of explosions derived from the seismic signals (Kanamori et al. 1984; Nishimura and Hamaguchi 1993; Chouet 2003; Zobin et al. 2009b).

The complex structure of the seismic signals associated with volcanic explosions as well as the problem of identification of the nature of

waveforms, composing these seismic signals, and unclear arrivals of the seismic phases make the study of the source process of explosion earthquakes difficult. To resolve the problem, many research scientists study the explosion event as a source that generates long-period and very-long-period seismic signals neglecting short-period components. The articles of Chouet (2019) and Kumagai (2019) describe in details this approach to quantify the volcanic explosions applying the waveform inversion and obtaining the characteristics of a centroid as the source of these vibrations. This approximation gives the integral characteristics of the source of explosion events based on a moment tensor and single-force representation of the source.

Another approach to quantify the volcanic explosions is based on the fine structure of broadband seismic signal as the information about the development of ascending magmatic process within the conduit before an explosion and during the explosion (Tameguri et al. 2002; Zobin et al. 2006, 2009b). The model proposed by Zobin et al. (2006, 2009b) describes the volcanic explosion process as consisting of the ascent of fragmented magma to the surface (stage 1) and a subsequent explosion (stage 2) (see Fig. 13). Based on this model, the energy of an explosion, E , may be calculated from the seismic signal corresponding to the co-explosion main phase.

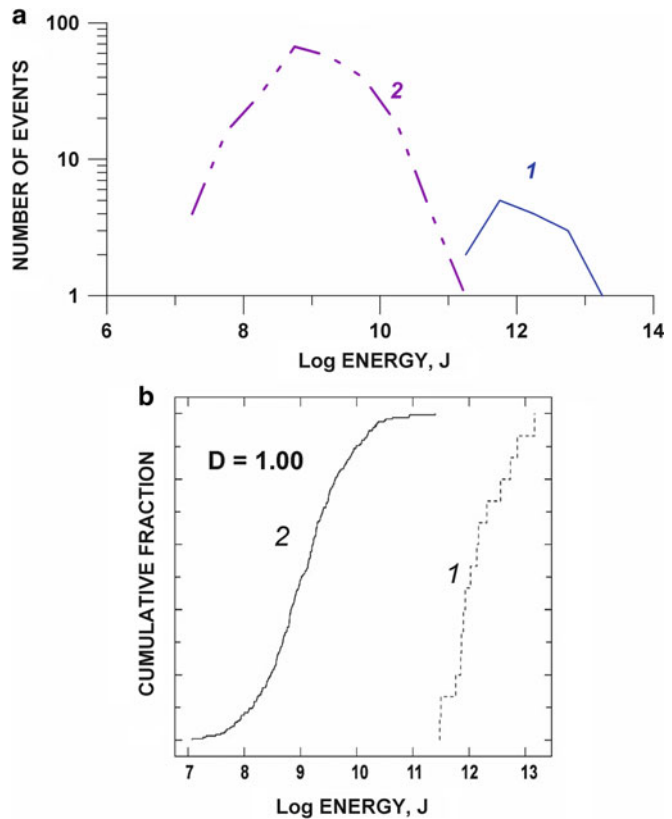
The energy values estimated for explosive eruptions occurring at Volcán de Colima during its 2004–2005 activity varied within a wide range, from 10^7 to 10^{13} J. The comparison of the energy-frequency distributions obtained for two databases consisting of 15 large explosions ($E > 10^{11}$ J) occurring during the March–September 2005 explosive sequence at Volcán de Colima and 236 small explosions ($E < 10^{11}$ J) of the November 2004 sequence at the same volcano (Fig. 26a) showed that the two curves obtained for these databases are separated at the energy size of 10^{11} J. Application of the Kolmogorov-Smirnov test (Fig. 26b) confirmed that these two curves characterize two independent samples of small and large explosions that are significant at the 0.01 critical level. This result demonstrated that the nature of the seismic signals, generated by large and small explosions, is different (Zobin 2017).

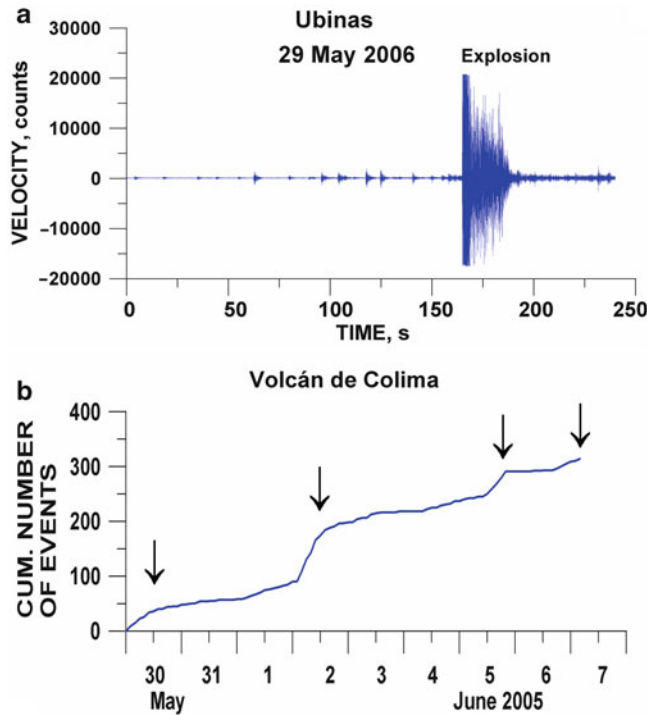
Microearthquakes

Swarms of microearthquakes have been observed during effusive and explosive activity at volcanoes of different types all over the world. Frequently they appear before the beginning of the lava extrusion or large explosion. Figure 27a shows the fragment of a sequence of microearthquakes preceding the 29 May 2006 explosion at Ubinas volcano, Peru. A sequence of large explosions, accompanied by the ash columns reaching to 7–8 km asl, occurred at Ubinas during May–October 2006. Forty-three of 134 explosions were preceded by the swarms of microearthquakes (Macedo et al. 2008). Similar effects of appearance of microearthquakes swarms before the large explosions were reported for the 2005 explosions at Volcán de Colima (Fig. 27b). The large explosions with the energy $E > 10^{11}$ J were preceded by a significant increase in the number

Volcano Seismology: An Introduction,

Fig. 26 Distribution in the energy for large explosions of the March–September 2005 sequence (1) and small explosions of the November 2004 sequence (2) at Volcán de Colima (a) and the Kolmogorov-Smirnov test comparison cumulative fraction plot for the sequences 1 and 2 (b). The statistic D is the maximum vertical deviation between two curves





Volcano Seismology: An Introduction, Fig. 27 (a) The sequence of long-period microearthquakes recorded before the large 29 May 2006 explosion at Ubinas volcano, Peru, at a distance of 2.5 km from the crater (short-period record, velocity, vertical component). The seismic records were provided by Orlando Macedo (Instituto Geofísico del

Peru). (b) Temporal variations in the cumulative number of microearthquakes recorded at a distance of 4 km from the crater (short-period record, velocity, vertical component) during the May–June 2005 explosive sequence at Volcán de Colima. The moments of explosions are shown by the arrows. (Modified from Zobin et al. 2010)

of microearthquakes (Zobin et al. 2010). The increase in the number of microearthquakes before the explosion appearance was noted also for the 2005 explosions at Mount St. Helens (Moran et al. 2008) and for the 5 April 2003 explosion at Stromboli (Carniel et al. 2006).

The swarms of microearthquakes preceded the 30 September 2004 lava extrusion at Volcán de Colima (Zobin 2017). A set of short-period seismograms recorded during 26 to 30 September 2004 is shown in Fig. 28. The rate of occurrence of microearthquakes slightly increased about 18 h after the beginning of sequence (Fig. 28b). The cumulative number continued to grow gradually until the beginning of the lava extrusion about 70 h after the beginning of the microearthquake sequence.

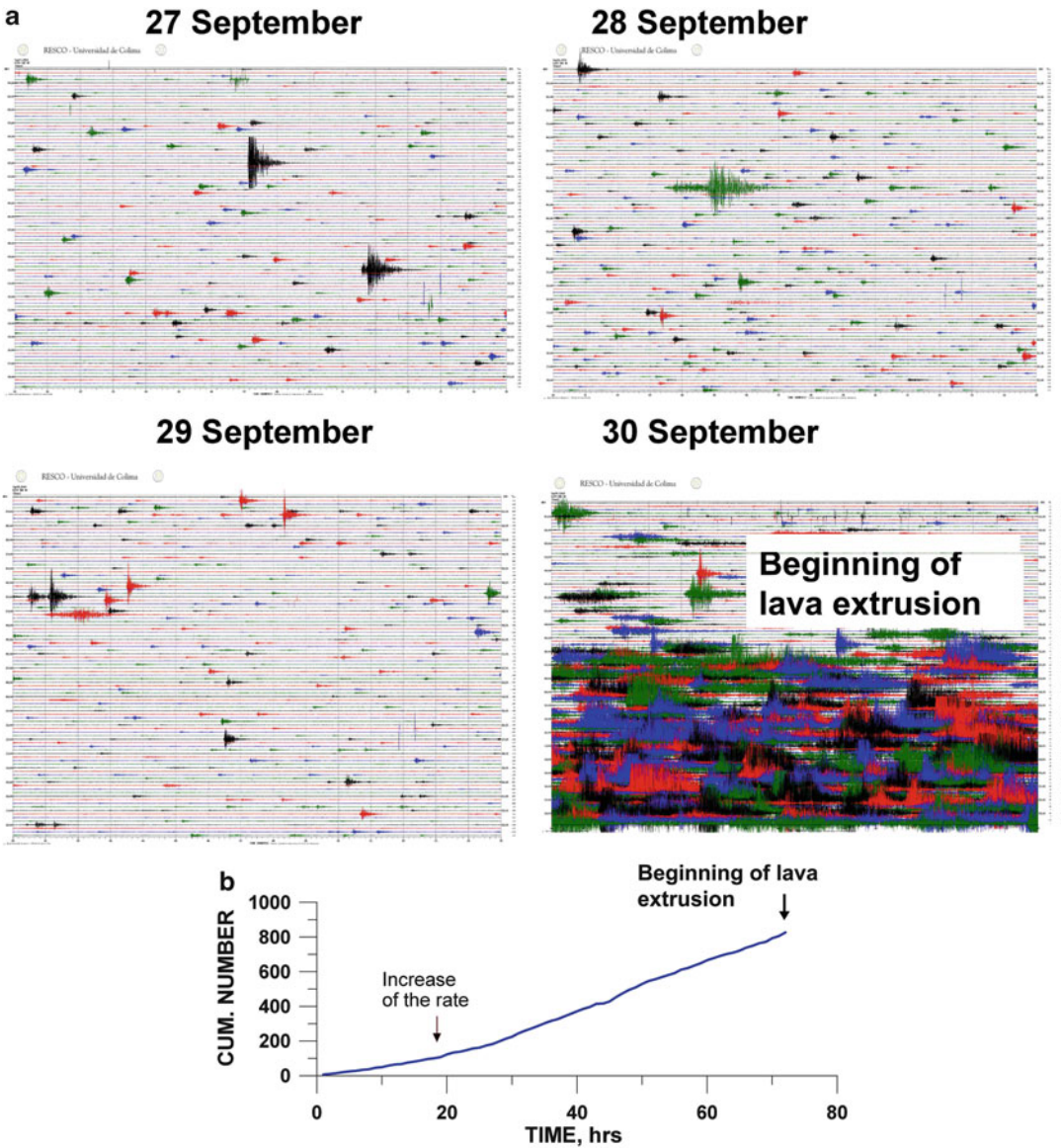
Significant effusive eruption interrupted the persistent explosive activity at Stromboli during

the period 27 February to 2 April 2007. Microearthquakes were recorded just before and after the 27 February effusive eruption. Before the onset of effusive eruption, they were of larger amplitudes (Martini et al. 2007).

Generally speaking, the seismic signals of different types recorded at volcanoes can serve as the eyes of volcanologists for monitoring and reconstruction of eruption activity.

Future Directions

As is seen from this short review, the quantitative theory of seismo-volcanic processes is being developed now only for some cases, and it is far from complete. Therefore, the future directions in development of *Volcano Seismology* may be of two types, experimental and theoretical. In the



Volcano Seismology: An Introduction, Fig. 28 Seismic records during 4 days of microearthquake activity at Volcán de Colima (short-period seismic station, vertical component 1.7 km from the crater) (a) and the curve of the cumulative number of microearthquakes

(b) before the 30 September 2004 lava extrusion. The beginning of extrusion is marked by a swarm of rockfall earthquakes on the seismogram of 30 September and is shown by an arrow on the curve. (Seismograms are courtesy of Colima Volcano Observatory)

experimental study of volcanic seismicity, we need in better identification of the nature of the seismic signals recorded at volcanoes. We use now the terms of “low-frequency” earthquake, “long-period” earthquake, “very-long-period” earthquake, volcanic tremor, or “hybrid” earthquake

referring only to the type of waveform but frequently without knowledge of the nature of generation of these signals. To resolve this problem, we need the dense networks of complex monitoring of volcanic activity, including simultaneous seismic, acoustic, geodetic, and video digital recording with

a high resolution of the signals and installed at distances at least from 1 to 10 km from the crater of active volcano. The networks have to be in operation for a long term, during a few cycles of the eruptions. This great collection of the good identified seismic signals of volcanic earthquakes may serve as a base for development of the quantitative theory of seismo-volcanic process.

Acknowledgments I thank my colleagues from all over the volcanic world for collaboration and useful discussions. The comments proposed by Robin Matoza, Gregory Waite, Willie Lee, and Robert Tilling helped me to improve the text. I use in my entry many seismic records and video images obtained at Colima Volcano Observatory by the teams headed by Gabriel Reyes and Mauricio Bretón, respectively. The catalogs and seismic records of volcanic signals were provided by my friends from different volcanological centers. The processing of the majority of digital signals was performed using the program DEGTRA provided by Mario Ordaz.

Bibliography

- Aki K (1992) State of the art in volcanic seismology. In: Gasparini P, Scarpa R, Aki K (eds) *Volcanic seismology*. Springer, Berlin/New York, pp 3–10
- Alparone S, Andronico D, Lodato L, Sgroi T (2003) Relationship between tremor and volcanic activity during Southeast Crater eruption on Mount Etna in early 2000. *J Geophys Res* 108:2241
- Arámbula-Mendoza R, Lesage P, Valdés-González C, Varley NR, Reyes-Dávila G, Navarro C (2011) Seismic activity that accompanied the effusive and explosive eruptions during the 2004–2005 period at Volcán de Colima México. *J Volcan Geotherm Res* 205:30–46
- Calder ES, Luckett R, Sparks RSJ, Voight B (2002) Mechanisms of lava dome instability and generation of rockfalls and pyroclastic flows at Soufrière Hills volcano, Montserrat. In: Druitt TH, Kokelaar BP (eds) *The eruption of Soufrière Hills volcano Montserrat from 1995 to 1999*, Geological society London memoirs, vol 21, pp 173–190
- Carniel R, Ortiz R, Di Cecca M (2006) Spectral and dynamical hints on the time scale of preparation of the 5 April 2003 explosion at Stromboli volcano. *Can J Earth Sci* 43:41–55
- Cashman KV, Sturtevant B, Papale P, Navon O (2000) Magmatic fragmentation. In: Sigurdsson H (ed) *Encyclopedia of volcanoes*. Academic, San Diego, pp 421–430
- Chouet B (1986) Dynamics of fluid-driven crack in three-dimensions by the finite difference method. *J Geophys Res* 91:13967–13992
- Chouet B (1996) New methods and future trends in seismological volcano monitoring. In: Scarpa R, Tilling R (eds) *Monitoring and mitigation of volcano hazards*. Springer, Berlin/New York, pp 23–98
- Chouet B (2003) Volcano seismology. *Pure Appl Geophys* 160:739–788
- Chouet B (2019) Volcanoes non-linear processes in. In: Meyers RA (ed) *Encyclopedia of complexity and systems science*, 2nd edn. Springer, Berlin
- Chouet B, Matoza RS (2013) A multi-decadal view of seismic methods for detecting precursors of magma movement and eruption. *J Volcanol Geotherm Res* 252:108–175
- Cosentino M, Gresta S, Lombardo G, Patané G, Riuscetti M, Schick R, Viglianisi A (1984) Features of volcanic tremors at Mt Etna (Sicily) during the Match-August 1983 eruption. *Bull Volcanol* 47:929–939
- Ferrucci F (1995) Seismic monitoring at active volcanoes. In: McGuire B et al (eds) *Monitoring active volcanoes*. UCL Press, London, pp 60–90
- Francis P (1998) *Volcanoes a planetary perspective*. Clarendon Press, Oxford, p 443
- Harlow DH, Power JA, Laguerta EP, Ambubuyog G, White RA, Hoblitt RP (1996) Precursory seismicity and forecasting of the June 15 1991 eruption of Mount Pinatubo. In: Newhall CG, Punongbayan RS (eds) *Fire and mud eruptions and lahars of Mount Pinatubo Philippines*. University of Washington Press, Seattle, pp 285–305
- JMA (2003) Report on the eruption of Usuzan volcano in 2000. Technical report of Japan meteorological 124. JMA, Tokyo, p 247 (In Japanese, English abstract)
- Julian B, Miller JB, Foulger GR (1998) Non-double-couple earthquakes I theory. *Rev Geophys* 36:525–549
- Kanamori H, Given JW, Lay T (1984) Analysis of seismic body waves excited by the Mount St Helens eruption of May 18 1980. *J Geophys Res* 89:1856–1866
- Kawakatsu H, Ohminato T, Ito H, Kuwahara Y, Kato T, Tsuruga K, Honda S, Yomogida K (1992) Broadband seismic observations at the Sakurajima volcano Japan. *Geoph Res Lett* 19:1959–1962
- Koyanagi R, Chouet B, Aki K (1987) Origin of volcanic tremor in Hawaii part I. In: Decker RW, Wright TLSPH (eds) *Volcanism in Hawaii*, Geological survey professional paper 1350, vol 2, pp 1221–1257
- Kumagai H (2019) Volcano seismic signals source quantification of. In: Meyers RA (ed) *Encyclopedia of complexity and systems science*, 2nd edn. Springer, Berlin
- Lahr JC, Chouet BA, Stephens CD, Power JA, Page RA (1994) Earthquake classification location and error analysis in a volcanic environment: implications for the magmatic system of the 1989–1990 eruptions at redoubt volcano Alaska. *J Volcanol Geotherm Res* 62:137–152
- Latter JH (1979) Volcanological observations at Tongariro national park types and classification of volcanic earthquakes 1976–1978, Report, vol 150. Geophysics Division, Wellington, p 60
- Macedo O, Métaixian J-P, Taipe E, Ramos D (2008) Actividad sismo-volcánica asociada a la erupción del volcán Ubinas en 2006–2008. In: XIII Congreso

- Latinoamericano de Geología 29 setiembre –3 octubre 2008. Lima, Perú
- Malone SD (1983) Volcanic earthquakes: examples from Mount St Helens. In: Kanamori H, Boschi E (eds) *Earthquakes: observations theory and interpretation*. North-Holland Publishing Company, Amsterdam, pp 436–455
- Martini M, Giudicepietro F, D'Auria L, Esposito AM, Caputo T, Curciotti R, De Cesare W, Orazi M, Scarpato G, Caputo A, Peluso R, Linde A, Sacks S (2007) Seismological monitoring of the February 2007 effusive eruption of the Stromboli volcano. *Ann Geofis* 50:775–788
- Minakami T (1960) Fundamental research for predicting volcanic eruptions. Part I. *Bull Earthq Res Inst Tokyo Univ* 38:497–544
- Molina I, Kumagai H, García-Aristizábal A, Nakano M, Mothes P (2008) Source process of very-long-period events accompanying long-period signals at Cotopaxi volcano, Ecuador. *J Volcanol Geotherm Res* 176:119–133
- Moran SC, Malone SD, Qamar AI, Thelen WA, Wright AK, Caplan-Auerbach J (2008) Seismicity associated with renewed dome building at Mount St. Helens 2004–2006. In: Sherrod DR, Scott WE, Stauffer PH (eds) *A volcano rekindled: the renewed eruption of Mount St. Helens 2004–2006*, US professional paper, vol 1750, Washington, pp 27–60
- Nettles M, Ekström G (1998) Faulting mechanism of anomalous earthquakes near Bárðunga volcano Iceland. *J Geophys Res* 103:17973–17983
- Newhall CG, Self S (1982) The volcanic explosivity index (VEI): an estimate of explosive magnitude for historical volcanism. *J Geophys Res (Oceans and Atmospheres)* 87:1231–1238
- Nishimura T, Hamaguchi H (1993) Scaling law of volcanic explosion earthquakes. *Geophys Res Lett* 20:2479–2482
- Okada H (2002) Report on the 2000 eruption of Usu and investigation of volcanic risk. Institute of Seismology and Volcanology, Sapporo, pp 3–9. In Japanese
- Omori F (1911–1913) The Usu-san eruption and earthquake and elevation phenomena. *Bull Imper Earthq Invest Com* 5:1–38
- Omori F (1912) The eruptions and earthquakes of the Asama-yama. *Bull Imper Earthq Invest Com* 6:1–147
- Omori F (1914) The Sakura-jima eruptions and earthquakes I. *Bull Imper Earthq Invest Com* 8:5–34
- Omori F (1922) The Sakura-jima eruptions and earthquakes VI. *Bull Imper Earthq Invest Com* 8:467–524
- Reyes-Dávila GA, De la Cruz-Reyna S (2002) Experience in the short-term eruption forecasting at Volcán de Colima Mexico and public response to forecasts. *J Volcanol Geotherm Res* 117:121–128
- Ripepe M, Ciliberto S, Della Schiava M (2001) Time constraints for modelling source dynamics of volcanic explosions at Stromboli. *J Geophys Res* 106:8713–8727
- Rodolfo KS, Umbal JV, Alonso RA, Remotigue CT, Paladio-Melosantos ML, Salvador JHG, Evangelista D, Millar Y (1996) Two years of lahars on the western flank of Mount Pinatubo: initiation flow processes deposits and attendant geomorphic and hydraulic changes. In: Newhall CG, Punongbayan RS (eds) *Fire and mud eruptions and lahars of Mount Pinatubo Philippines*. University of Washington Press, Seattle, pp 989–1014
- Rogers N, Hawkesworth C (2000) Composition of magmas. In: Sigurdsson H (ed) *Encyclopedia of volcanoes*. Academic, San Diego, pp 115–132
- Rubin AM, Gillard D, Got J-L (1998) A reinterpretation of seismicity associated with the January 1983 dike intrusion at Kilauea volcano Hawaii. *J Geophys Res* 103:10003–10015
- Rutherford MJ, Gardner JE (2000) Rates of magma ascent. In: Sigurdsson H (ed) *Encyclopedia of volcanoes*. Academic, San Diego, pp 207–217
- Saraó A, Panza GF, Privitera E, Cocina O (2001) Non-double-couple mechanisms in the seismicity preceding the 1991–1992 Etna volcano eruption. *Geophys J Int* 145:319–335
- Schick R, Lombardo G, Patané G (1982) Volcanic tremors and shocks associated with eruptions at Etna (Sicily) September 1980. *J Volcanol Geotherm Res* 14:261–279
- Shuler A, Nettles M, Ekström G (2013a) Global observation of vertical-CLVD earthquakes at active volcanoes. *J Geophys Res* 118:138–164
- Shuler A, Ekström G, Nettles M (2013b) Physical mechanisms for vertical-CLVD earthquakes at active volcanoes. *J Geophys Res* 118:1–18
- Takahashi T (2007) *Debris flow*. Taylor & Francis, London, p 448
- Takahashi T, Satofuka Y (2002) Generalized theory of stony and turbulent muddy debris flow and its practical model. *J Jap Soc ECE* 55:33–42
- Takahashi T, Tsujimoto H (2000) A mechanical model for Merapi-type pyroclastic flow. *J Volcanol Geotherm Res* 98:91–115
- Tameguri T, Iguchi M, Ishihara K (2002) Mechanism of explosive eruptions from moment tensor analyses of explosion earthquakes at Sakurajima volcano Japan. *Bull Volcanol Soc Japan* 47:197–215
- Tokarev PI (1966) Eruptions and seismic regime of volcanoes of Klyuchevskaya Group. Nauka, Moscow, p 118. in Russian
- Tokarev PI (1978) Prediction and characteristics of the 1975 eruption of Tolbachik volcano Kamchatka. *Bull Volcanol* 41:1–8
- Tuffen H, Smith R, Sammonds PR (2008) Evidence for seismogenic fracture of silicic magma. *Nature* 453:511–514
- Uhira K, Yamasato H, Takeo M (1994) Source mechanism of seismic waves excited by pyroclastic flows observed at Unzen volcano Japan. *J Geophys Res* 99:17757–17773
- Uhira K, Baba T, Mori H, Katayama H, Hamada N (2005) Earthquake swarms preceding the 2000 eruption of Miyakejima volcano, Japan. *Bull Volcanol* 67:219–230
- Waite GP, Chouet BA, Dawson PB (2008) Eruption dynamics at Mount St Helens imaged from broadband seismic waveforms: interaction of the shallow

- magmatic and hydrothermal systems. *J Geophys Res* 113:B02305
- Wood HO (1913) Hawaiian volcano observatory. *Bull Seismol Soc Am* 3:14–19
- Yamasato H (1997) Quantitative analysis of pyroclastic flows using infrasonic and seismic data at Unzen volcano Japan. *J Phys Earth* 45:397–416
- Yokoyama I, Seino M (2000) Geophysical comparison of the three eruptions in the 20th century of Usu volcano, Japan. *Earth Planet Space* 52:73–89
- Zobin VM (1992) Volcanic earthquakes and their role in volcano-tectonic process. *Volcanol Seismol* 5-6:87–100. in Russian
- Zobin VM (2017) Introduction to volcanic seismology, 3rd edn. Elsevier, Amsterdam/New York/Tokyo, p 559
- Zobin VM, Nishimura Y, Miyamura J (2005a) The nature of volcanic earthquake swarm preceding the 2000 flank eruption at Usu volcano Hokkaido Japan. *Geophys J Int* 163:265–275
- Zobin VM, Orozco-Rojas J, Reyes-Dávila GA, Navarro C (2005b) Seismicity of an andesitic volcano during block-lava effusion: Volcán de Colima México November 1998–January 1999. *Bull Volcanol* 67:679–688
- Zobin VM, Navarro C, Reyes-Dávila G, Orozco J, Bretón M, Tellez A, Reyes-Alfaro G, Vázquez H (2006) The methodology of quantification of volcanic explosions from broadband seismic signals and its application to the 2004–2005 explosions at Volcán de Colima México. *Geophys J Int* 167:467–478
- Zobin VM, Varley NR, González M, Orozco J, Reyes-Dávila G, Navarro C, Bretón M (2008a) Monitoring the 2004 andesitic block-lava extrusion at Volcán de Colima México from seismic activity and SO₂ emission. *J Volcanol Geoth Res* 177:367–377
- Zobin VM, Reyes GA, Guevara E, Bretón M (2008b) Seismological constraints on the position of the fragmentation surfaces in the volcano conduit. *Earth Planet Sci Lett* 275:337–341
- Zobin VM, Plascencia I, Reyes G, Navarro C (2009a) The characteristics of seismic signals produced by lahars and pyroclastic flows: Volcán de Colima México. *J Volcanol Geotherm Res* 179:157–167
- Zobin VM, Reyes GA, Guevara E, Bretón M (2009b) Scaling relationship for Vulcanian explosions derived from broadband seismic signals. *J Geophys Res* 114: B03203
- Zobin VM, Melnik OE, Gonzalez M, Macedo O, Bretón M (2010) Swarms of micro-earthquakes associated with the 2005 Vulcanian explosion sequence at Volcán de Colima México. *Geophys J Int* 182:808–828



Source Quantification of Volcanic-Seismic Signals

Hiroyuki Kumagai
Graduate School of Environmental Studies,
Nagoya University, Nagoya, Japan

Article Outline

Glossary
Definition of the Subject
Introduction
Phenomenological Representation of Seismic Sources
Waveform Inversion
Spectral Analysis
Fluid-Solid Interactions
Future Directions
Green's Functions
Moment Tensor for a Spherical Source
Moment Tensor for a Cylindrical Source
Bibliography

Glossary

Autoregressive equation A difference form of the equation of motion of a linear dynamic system, which is a basic equation to determine the complex frequencies (frequencies and Q factors) of decaying harmonic oscillations in observed signals.

Crack wave A dispersive wave generated by fluid-solid interactions in a crack. The phase velocity of the crack wave is smaller than the acoustic velocity of the fluid in the crack.

Moment tensor A point seismic source representation defined by the first-order moment of the equivalent body force or the stress glut. Slip on a fault as well as volumetric changes such as an isotropic expansion and tensile crack can be represented by the moment tensor.

Waveform inversion An approach to estimate source mechanisms and locations of seismic

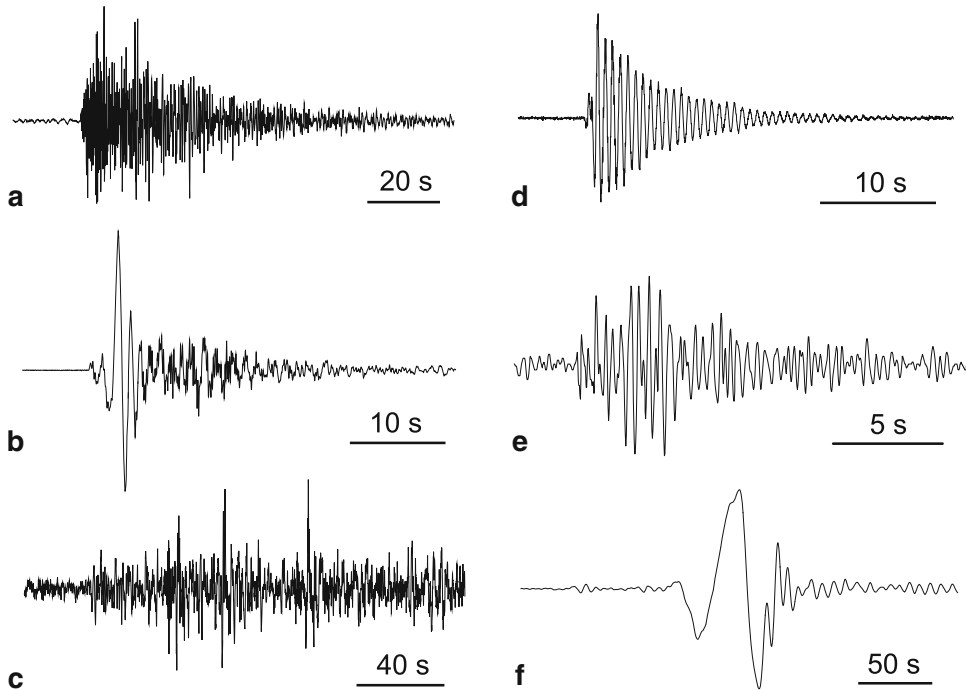
events by finding the best fits between observed and synthesized seismograms.

Definition of the Subject

Volcano seismicity produces a wide variety of seismic signals that provide glimpses of the internal dynamics of volcanic systems. Quantitative approaches to analyze and interpret volcano-seismic signals have been developed since the late 1970s. The availability of seismic equipments with wide frequency and dynamic ranges since the early 1990s revealed a further variety of volcano-seismic signals in oscillation periods longer than a few seconds. Quantification of the sources of volcano-seismic signals is crucial to achieving a better understanding of the physical states and dynamics of magmatic and hydrothermal systems.

Introduction

Volcanoes occur in tectonically active regions of the earth where magmatic and hydrothermal fluids display complex interactions with volcanic rocks and the atmosphere. Volcano seismicity is the manifestation of such complex interactions occurring inside of volcanic edifices. Volcano seismology aims at achieving a better understanding of the physical states and dynamics of magmatic and hydrothermal systems beneath active volcanoes through observation, analysis, and interpretation of volcano-seismic signals. It is well known that volcano-seismic signals have a wide variety of signatures (Fig. 1), which have been classified by various schemes (e.g., Chouet 1996a; Kawakatsu and Yamamoto 2007; McNutt 1996; Minakami 1974). The volcano-tectonic (VT) earthquake, low-frequency (LF) or long-period (LP) event, tremor, very-long-period (VLP) event, hybrid event, and explosion earthquake have been traditionally used to specify volcano-seismic signals. Here, LF denotes frequencies between 0.5 and



Source Quantification of Volcanic-Seismic Signals, Fig. 1 Various volcano-seismic signals: (a) volcano-tectonic earthquake that occurred between Miyakejima and Kozujima, Japan; (b) explosion earthquake observed at Asama volcano, Japan; (c) tremor that occurred beneath

Mt. Fuji, Japan; (d) long-period event observed at Kusatsu-Shirane volcano, Japan; (e) long-period event observed at Guagua Pichincha volcano, Ecuador; (f) very-long-period event that occurred beneath Miyakejima, Japan (low-pass filtered at 20 s)

5 Hz (Kawakatsu and Yamamoto 2007), and LP and VLP denote periods between 0.2 and 2 s and longer than a few seconds, respectively (Chouet 1996a).

In contrast to the various classification schemes, source processes of volcano-seismic signals may be classified into three main families: (i) brittle failure in volcanic rock triggered by fluid movement, (ii) transient pressure disturbance caused by mass transport and/or volumetric change, and (iii) resonance of a fluid-filled resonator excited by a pressure disturbance. These three processes may be linked to the signals as follows: (i) VT earthquake; (ii) VLP event and explosion earthquake; (iii) LF or LP event, tremor, and VLP event; and (i) + (iii) hybrid event. VT earthquakes, sometimes called high-frequency (HF) earthquakes, are indistinguishable from ordinary tectonic earthquakes but occur in volcanic regions due to process (i). LF or LP events have signals characterized by decaying

oscillations, whereas the signature of tremor shows continuous oscillations. These oscillatory signatures can be interpreted as process (iii), in which the excitations are impulsive for LF or LP events and successive for tremor. Most VLP events are characterized by impulsive signatures, which may be generated by process (ii). However, some VLP events show sustained oscillations or decaying harmonic oscillations, which may reflect process (iii). Hybrid events show mixed characters of both VT earthquakes and LP events, suggesting that processes (i) and (iii) both take place at the source. Explosion earthquakes occur in association with eruptions in which process (ii) is dominant.

Recently, there have been remarkable advances in quantitative understanding of the sources of volcano-seismic signals. These advances were made possible by the development of source models and analysis techniques coupled with increased computer capacity and the availability

of seismic observation equipment with wide frequency and dynamic ranges. Major studies that have contributed to these advances are summarized as follows: detailed magma intrusion processes imaged by VT earthquakes (e.g., Brandsdottir and Einarsson 1979; Fukuyama et al. 2001; Hayashi and Morita 2003; Morita et al. 2006; Rubin and Gillard 1998; Rubin et al. 1998; Toda et al. 2002); quantitative interpretations of oscillations in LP and VLP events and tremor to diagnose the states of fluids at the sources of these signals (e.g., Aki et al. 1977; Aoyama and Takeo 2001; Chouet 1985, 1986, 1988, 1992; Chouet and Julian 1985; Chouet et al. 1994; De Angelis and McNutt 2005; Ferrazzini and Aki 1987; Fujita and Ida 2003; Fujita et al. 1995, 2004; Gil Cruz and Chouet 1997; Jousset et al. 2003; Julian 1994; Kumagai 2006; Kumagai and Chouet 1999, 2000; Kumagai et al. 2002a; Kumagai et al. 2002b, 2005; Lesage et al. 2002; Lin et al. 2005; Molina et al. 2004; Morrissey and Chouet 1997, 2001; Nakamichi et al. 2003; Nakano and Kumagai 2005a, b; Nakano et al. 1998, 2003, 2007; Neuberg et al. 2000; Yamamura and Kawakatsu 1998); tracking the sources of LP events and tremor (e.g., Almendros et al. 2001a, b, 2002a; Battaglia and Aki 2003; Chouet et al. 1997; Goldstein and Chouet *n.d.*; Métaxian et al. 1997; Saccorotti and Del Pezzo 2000; Saccorotti et al. 2001); and waveform analysis of impulsive and oscillatory signatures in VLP events and explosion earthquakes to investigate source dynamics (e.g., Almendros et al. 2002b; Arciniega-Ceballos et al. 1999, 2003; Aster et al. 2003; Auger et al. 2006; Chouet et al. 1999, 2003, 2005, 2006; Dawson et al. 1998; Hidayat et al. 2000, 2002; Hill et al. 2002; Iguchi 1994; Julian et al. 1998; Kanamori and Given 1982; Kanamori et al. 1984; Kaneshima et al. 1996; Kawakatsu et al. 1992, 1994, 2000; Kobayashi et al. 2003; Kumagai et al. 2001, 2003; Legrand et al. 2000; Neuberg et al. 1994; Nishimura et al. 2000, 2002; Ohminato 2006; Ohminato et al. 1998, 2006; Rowe et al. 1998; Takeo et al. 1990; Tameguri et al. 2002; Uhira and Takeo 1994; Yamamoto et al. 1999, 2002). These volcano seismological studies have been reviewed by various authors

(Chouet 1996a, b, 2003; Kawakatsu and Yamamoto 2007; McNutt 1996; McNutt 2005; Neuberg 2000).

The purpose of this article is to provide a systematic presentation of the theoretical basis for quantification of the sources of volcano-seismic signals suitable for readers, including graduate students and young researchers, who are new to the study of volcano seismology. I focus on four subjects: (1) phenomenological representation of seismic sources, (2) waveform inversion to estimate source mechanisms, (3) spectral analysis based on an autoregressive model, and (4) physical properties of fluid-solid coupled waves. Some additional notes are provided in Appendixes. These constitute the basic elements needed to quantify the sources of volcano-seismic signals. Applications of quantitative approaches are not fully reviewed in this manuscript. Readers are encouraged to consult both the original research papers and the review papers cited above.

Phenomenological Representation of Seismic Sources

We first derive the phenomenological representation of seismic sources using single force and moment tensor. This representation provides the basis to study source processes of volcano seismicity. I follow the approach presented by Backus and Mulcahy (1976) (hereafter referred to as BM) and Aki and Richards (2002) for the moment tensor representation. The theory elaborated by Takei and Kumazawa (1994, 1995) is used to introduce single force in seismic sources.

Stress Glut

We consider indigenous seismic sources occurring within the closed earth system. This means that external bodies and detached mass are excluded. Let us assume an isotropic medium for the earth, which is represented by density ρ and Lamé's constants λ and μ . The exact equation of motion of the earth is given by

$$\rho \frac{\partial^2 u_i}{\partial t^2} = \frac{\partial \sigma_{ni}^{\text{true}}}{\partial x_n}, \quad (1)$$

where u_i and $\sigma_{ni}^{\text{true}}$ are the true displacement field and true stress tensor, respectively, actually produced in the earth. These are functions of space $\mathbf{x} = (x_1, x_2, x_3)$ and time t . The summation convention for repeated subscripts is used throughout this paper unless otherwise stated. In BM's theory, the failure of the elastic stress-strain relationship in source processes, such as slip across a plane that cannot be described by the linear elastodynamics, is represented by an equivalent body force. We use a model stress $\sigma_{ni}^{\text{model}}$ given by the elastic stress-strain relationship as

$$\sigma_{ni}^{\text{model}} = \lambda \frac{\partial u_k}{\partial x_k} \delta_{ni} + \mu \left(\frac{\partial u_n}{\partial x_i} + \frac{\partial u_i}{\partial x_n} \right), \quad (2)$$

where δ_{ni} is the Kronecker symbol ($\delta_{ni} = 0$ for $n \neq i$ and $\delta_{ni} = 1$ for $n = i$). By replacing the true stress by the model stress, Eq. (1) can be rewritten as follows:

$$\rho \frac{\partial^2 u_i}{\partial t^2} = \frac{\partial \sigma_{ni}^{\text{model}}}{\partial x_n} + f_i^S, \quad (3)$$

where f_i^S is the equivalent body force defined as

$$f_i^S = \frac{\partial}{\partial x_n} (\sigma_{ni}^{\text{true}} - \sigma_{ni}^{\text{model}}). \quad (4)$$

The difference between $\sigma_{ni}^{\text{true}}$ and $\sigma_{ni}^{\text{model}}$ is called the stress glut. The equivalent body force, which is the spatial derivative of the stress glut, describes

the source and vanishes outside the source region for motion in a completely elastic medium.

We introduce Green's function $G_{ij}(\mathbf{x}, t; \boldsymbol{\eta}, \tau)$, which is the i th component of displacement at $(\mathbf{x}, t)$ excited by the unit impulse applied at $\mathbf{x} = \boldsymbol{\eta}$ and $t = \tau$ in the j -direction. $G_{ij}(\mathbf{x}, t; \boldsymbol{\eta}, \tau)$ satisfies the following equation:

$$\rho \frac{\partial^2 G_{ij}}{\partial t^2} = \frac{\partial}{\partial x_n} \left\{ \lambda \frac{\partial G_{kj}}{\partial x_k} \delta_{ni} + \mu \left(\frac{\partial G_{nj}}{\partial x_i} + \frac{\partial G_{ij}}{\partial x_n} \right) \right\} + \delta(\mathbf{x} - \boldsymbol{\eta}) \delta(t - \tau) \delta_{ij}. \quad (5)$$

Note that $G_{ij}(\mathbf{x}, t; \boldsymbol{\eta}, \tau) = G_{ij}(\mathbf{x}, t - \tau; \boldsymbol{\eta}, 0)$. The displacement u_i can then be described by using the equivalent body force and Green's functions as follows ("Appendix A: Green's Functions"):

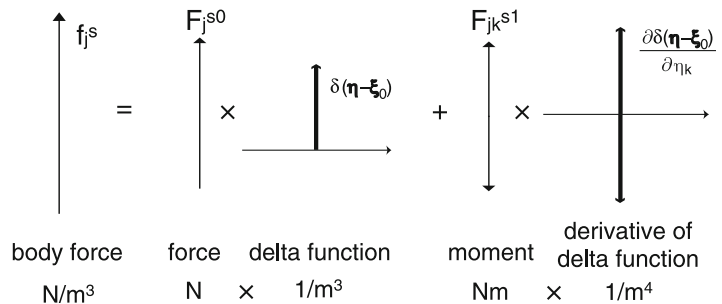
$$u_i(\mathbf{x}, t) = \int_{-\infty}^{\infty} \int_V f_j^S(\boldsymbol{\eta}, \tau) G_{ij}(\mathbf{x}, t - \tau; \boldsymbol{\eta}, 0) dV(\boldsymbol{\eta}) d\tau. \quad (6)$$

BM expanded f_j^S in terms of polynomial moments at a particular point $\boldsymbol{\xi}_0$ as

$$\begin{aligned} f_j^S(\boldsymbol{\eta}, \tau) &= F_j^{S0}(\boldsymbol{\xi}_0, \tau) \delta(\boldsymbol{\eta} - \boldsymbol{\xi}_0) \\ &+ F_{jk}^{S1}(\boldsymbol{\xi}_0, \tau) \frac{\partial \delta(\boldsymbol{\eta} - \boldsymbol{\xi}_0)}{\partial \eta_k} + \dots \\ &= f_j^{S0}(\boldsymbol{\eta}, \tau) + f_j^{S1}(\boldsymbol{\eta}, \tau) + \dots \end{aligned} \quad (7)$$

This expansion is schematically shown in Fig. 2, which may help the reader to understand that F_j^{S0} and F_{jk}^{S1} have the dimensions of force and moment, respectively.

Source Quantification of Volcanic-Seismic Signals, Fig. 2 Schematic diagram of the expansion of the equivalent body force f_j^S in terms of polynomial moments



Single Force

F_j^{S0} is given as

$$F_j^{S0}(\xi_0, \tau) = \iiint_V f_j^S(\eta, \tau) dV(\eta), \quad (8)$$

which is the total force. This must be zero because of the conservation of linear momentum (no force appears or disappears in the earth). Therefore,

$$\iiint_V f_j^S(\eta, \tau) dV(\eta) = 0. \quad (9)$$

This states that no single force exists in any indigenous seismic sources in the scheme presented by BM. However, we will see later that a single force exists in indigenous sources if the nonlinear effect of mass advection is taken into account.

Moment Tensor

We next examine the equivalent body force arising from the first-order moment

$$f_j^{S1}(\eta, \tau) = F_{jk}^{S1}(\xi_0, \tau) \frac{\partial \delta(\eta - \xi_0)}{\partial \eta_k}. \quad (10)$$

If we consider a displacement discontinuity across a surface Σ , f_j^{S1} may be expressed as (Aki and Richards 2002)

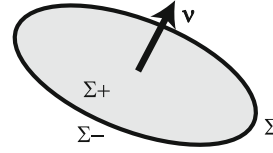
$$f_j^{S1}(\eta, \tau) = - \iint_{\Sigma} m_{jk}(\xi, \tau) \frac{\partial}{\partial \eta_k} \delta(\eta - \xi) d\Sigma(\xi), \quad (11)$$

where m_{jk} is the moment density tensor given as

$$m_{jk} = \lambda v_l [u_l] \delta_{jk} + \mu (v_j [u_k] + v_k [u_j]). \quad (12)$$

Here, $[\mathbf{u}]$ is the displacement discontinuity defined by $[\mathbf{u}] = \mathbf{u}|_{\Sigma_+} - \mathbf{u}|_{\Sigma_-}$, and $\boldsymbol{\nu} = (v_1, v_2, v_3)$ is the unit vector normal to Σ , where $\mathbf{u}|_{\Sigma_+}$ and $\mathbf{u}|_{\Sigma_-}$ are displacements on the Σ_+ and the Σ_- sides of Σ , respectively (see Fig. 3). An explicit expression of m_{jk} is

$$\mathbf{m} = \begin{pmatrix} \lambda v_k [u_k] + 2\mu v_1 [u_1] & \mu (v_1 [u_2] + v_2 [u_1]) \\ \mu (v_2 [u_1] + v_1 [u_2]) & \lambda v_k [u_k] + 2\mu v_2 [u_2] \\ \mu (v_3 [u_1] + v_1 [u_3]) & \mu (v_3 [u_2] + v_2 [u_3]) \\ \mu (v_1 [u_3] + v_3 [u_1]) & \\ \mu (v_2 [u_3] + v_3 [u_2]) & \\ \lambda v_k [u_k] + 2\mu v_3 [u_3] & \end{pmatrix} \quad (13)$$



Source Quantification of Volcanic-Seismic Signals, Fig. 3

An internal surface Σ across which a displacement discontinuity occurs. The displacement discontinuity is denoted by $[\mathbf{u}] = \mathbf{u}|_{\Sigma_+} - \mathbf{u}|_{\Sigma_-}$, where $\mathbf{u}|_{\Sigma_+}$ and $\mathbf{u}|_{\Sigma_-}$ are displacements on the Σ_+ and Σ_- sides of Σ , respectively (Aki and Richards 2002)

Therefore $m_{jk} = m_{kj}$. Using Green's functions, the displacement field due to f_j^{S1} is expressed as

$$\begin{aligned} u_i^{S1}(\mathbf{x}, t) &= \int_{-\infty}^{\infty} \iiint_V f_j^{S1}(\eta, \tau) G_{ij}(\mathbf{x}, t - \tau; \eta, 0) dV(\eta) d\tau \\ &= \int_{-\infty}^{\infty} \iiint_V \left[- \iint_{\Sigma} m_{jk}(\xi, \tau) \frac{\partial}{\partial \eta_k} \delta(\eta - \xi) d\Sigma(\xi) \right] \\ &\quad \cdot G_{ij}(\mathbf{x}, t - \tau; \eta, 0) dV(\eta) d\tau \\ &= \int_{-\infty}^{\infty} \iint_{\Sigma} \left[m_{jk}(\xi, \tau) \iiint_V - \frac{\partial}{\partial \eta_k} \delta(\eta - \xi) G_{ij}(\mathbf{x}, t - \tau; \eta, 0) dV(\eta) \right] d\Sigma(\xi) d\tau \\ &= \int_{-\infty}^{\infty} \iint_{\Sigma} m_{jk}(\xi, \tau) \frac{\partial}{\partial \xi_k} G_{ij}(\mathbf{x}, t - \tau; \xi, 0) d\Sigma(\xi) d\tau, \end{aligned} \quad (14)$$

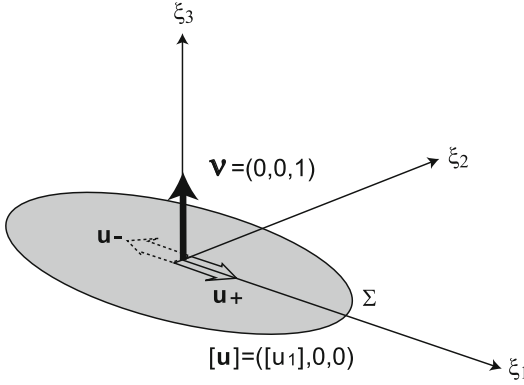
where I used the following property:

$$\begin{aligned} &\iiint_V \frac{\partial}{\partial \eta_k} \delta(\eta - \xi) G_{ij}(\mathbf{x}, t - \tau; \eta, 0) dV(\eta) \\ &= - \frac{\partial}{\partial \xi_k} G_{ij}(\mathbf{x}, t - \tau; \xi, 0). \end{aligned} \quad (15)$$

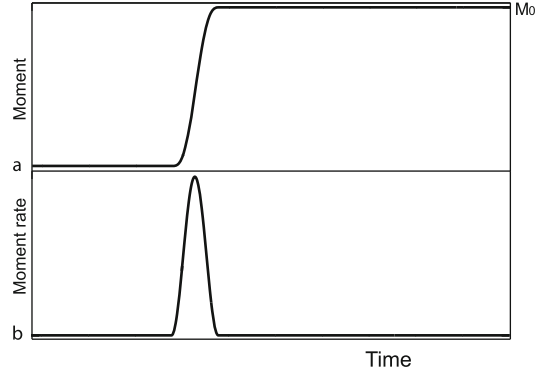
We assume that the source is small compared to the wavelengths of seismic waves and approximate the source as a point at ξ_0 . Then, Eq. (14) can be written as

$$u_i^{S1}(\mathbf{x}, t) = \int_{-\infty}^{\infty} M_{jk}(\xi_0, \tau) \frac{\partial}{\partial \xi_k} G_{ij}(\mathbf{x}, t - \tau; \xi_0, 0) d\tau, \quad (16)$$

where M_{jk} is the moment tensor defined as



Source Quantification of Volcanic-Seismic Signals, Fig. 4 Slip $[u_1]$ in the plane $\xi_3 = 0$



Source Quantification of Volcanic-Seismic Signals, Fig. 5 Moment and moment-rate functions for slip across a plane in the form of a steplike function

$$M_{jk}(\xi_0, \tau) = \iint_{\Sigma} m_{jk}(\xi, \tau) d\Sigma(\xi). \quad (17)$$

Slip on a Fault

We consider slip $[u_1]$ in the plane $\xi_3 = 0$ (see Fig. 4). In this case, $\nu = (0, 0, 1)$ and $[u] = ([u_1], 0, 0)$. Substituting ν and $[u]$ into Eq. (13), we obtain the moment density tensor m_{jk} as.

$$\mathbf{m} = \begin{pmatrix} 0 & 0 & \mu[u_1(\xi, \tau)] \\ 0 & 0 & 0 \\ \mu[u_1(\xi, \tau)] & 0 & 0 \end{pmatrix}, \quad (18)$$

and the moment tensor M_{jk} as

$$\mathbf{M} = M(\xi_0, \tau) \begin{pmatrix} 0 & 0 & 1 \\ 0 & 0 & 0 \\ 1 & 0 & 0 \end{pmatrix}, \quad (19)$$

where

$$M(\xi_0, \tau) = \mu \iint_{\Sigma} [u_1(\xi, \tau)] d\Sigma(\xi). \quad (20)$$

This is called the moment function or source-time function. The time history of slip across a plane is typically represented by a steplike function, and therefore $M(\xi_0, \tau)$ displays a steplike function (Fig. 5a). We can also define the time derivative of M as

$$\dot{M}(\xi_0, \tau) = \mu \frac{\partial}{\partial \tau} \iint_{\Sigma} [u_1(\xi, \tau)] d\Sigma(\xi). \quad (21)$$

This is called the moment-rate function, which is an impulsive function (Fig. 5b). After slip stops, M has a finite value M_0 (Fig. 5a), which is given as

$$M_0 = \mu A \overline{u_1}. \quad (22)$$

A and $\overline{u_1}$ are the slip area and average slip, respectively, which are defined as

$$A = \iint_{\Sigma} d\Sigma, \quad (23)$$

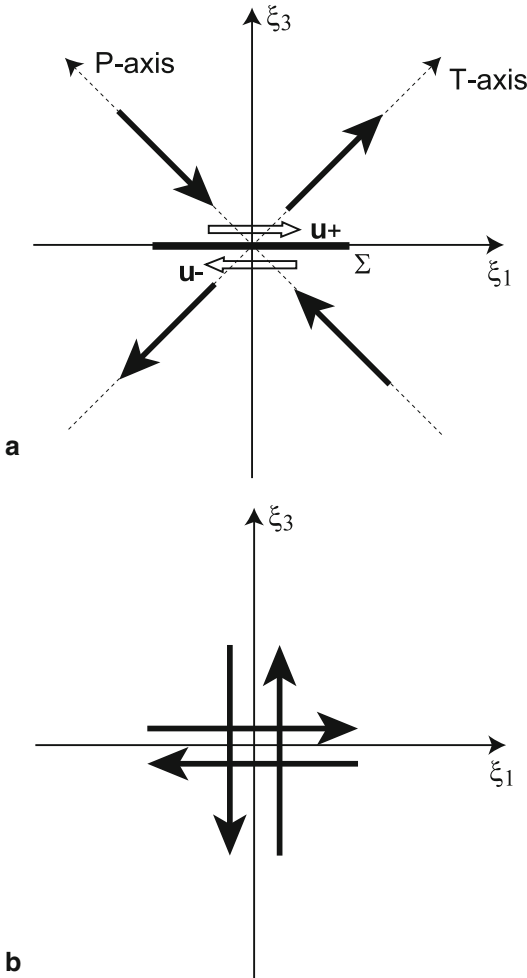
$$\overline{u_1} = \frac{\iint_{\Sigma} [u_1(\xi, \tau = \infty)] d\Sigma}{A}, \quad (24)$$

where $\tau = \infty$ represents time after the termination of slip. M_0 is called the seismic moment. When $\tau = \infty$, M_{jk} becomes

$$\mathbf{M} = M_0 \begin{pmatrix} 0 & 0 & 1 \\ 0 & 0 & 0 \\ 1 & 0 & 0 \end{pmatrix}. \quad (25)$$

This matrix can be diagonalized to

$$\mathbf{M}' = M_0 \begin{pmatrix} 1 & 0 & 0 \\ 0 & 0 & 0 \\ 0 & 0 & -1 \end{pmatrix}, \quad (26)$$

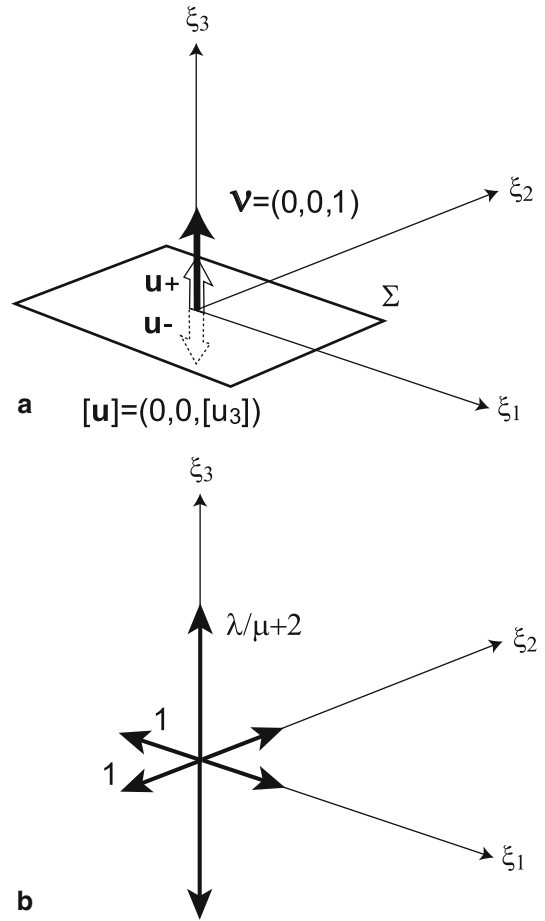


Source Quantification of Volcanic-Seismic Signals, Fig. 6 (a) Force system for slip $[u_1]$ in the plane $\xi_3 = 0$. (b) A double couple equivalent to the force system in Fig. 6a

where three eigenvalues (1, 0, and -1) correspond to the T (maximum compression), B (null), and P (minimum compression) axes, respectively. As shown in Fig. 6, this system constitutes a double couple, which is equivalent to fault slip.

Tensile Crack

We examine a crack such that slip $[u_3]$ is nonzero in the plane $\xi_3 = 0$ (a horizontal crack; see Fig. 7). In this case, $\mathbf{v} = (0, 0, 1)$ and $[\mathbf{u}] = (0, 0, [u_3])$. Therefore, we find from Eq. (13) that



Source Quantification of Volcanic-Seismic Signals, Fig. 7 (a) A tensile horizontal crack in which slip $[u_3]$ is nonzero in the plane ξ_3 . (b) Force system equivalent to a tensile horizontal crack

$$\mathbf{m} = \begin{pmatrix} \lambda[u_3(\xi, \tau)] & 0 & 0 \\ 0 & \lambda[u_3(\xi, \tau)] & 0 \\ 0 & 0 & (\lambda + 2\mu)[u_3(\xi, \tau)] \end{pmatrix}. \quad (27)$$

The moment tensor for a horizontal crack is given as

$$\mathbf{M} = M(\xi_0, \tau) \begin{pmatrix} \lambda/\mu & 0 & 0 \\ 0 & \lambda/\mu & 0 \\ 0 & 0 & \lambda/\mu + 2 \end{pmatrix}, \quad (28)$$

and

$$\begin{aligned}
 M(\xi_0, \tau) &= \mu \iint_{\Sigma} [u_3(\xi, \tau)] d\Sigma(\xi) \\
 &= \mu A \overline{u_3}(\tau) \\
 &= \mu \Delta V(\tau),
 \end{aligned} \quad (29)$$

where ΔV represents the incremental volume change of the crack.

$$\mathbf{M} = \mu \Delta V \begin{pmatrix} \lambda/\mu + 2 \sin^2 \theta \cos^2 \phi & 2 \sin^2 \theta \sin \phi \cos \phi & 2 \sin \theta \cos \theta \cos \phi \\ 2 \sin^2 \theta \sin \phi \cos \phi & \lambda/\mu + 2 \sin^2 \theta \sin^2 \phi & 2 \sin \theta \cos \theta \sin \phi \\ 2 \sin \theta \cos \theta \cos \phi & 2 \sin \theta \cos \theta \sin \phi & \lambda/\mu + 2 \cos^2 \theta \end{pmatrix}, \quad (30)$$

where $\Delta V = \int \int_{\Sigma} [u] d\Sigma$.

For a vertical crack ($\theta = \pi/2$),

$$\mathbf{M} = \mu \Delta V \begin{pmatrix} \lambda/\mu + 2 \cos^2 \phi & 2 \sin \phi \cos \phi & 0 \\ 2 \sin \phi \cos \phi & \lambda/\mu + 2 \sin^2 \phi & 0 \\ 0 & 0 & \lambda/\mu \end{pmatrix}. \quad (31)$$

A tensile crack is equivalent to a superposition of three vector dipoles with amplitude ratios 1:1: ($\lambda + 2\mu$)/ λ (Fig. 7). If $\lambda = \mu$, the amplitude ratios are represented by 1:1:3. If $\lambda = 2\mu$, which may be appropriate either for volcanic rock near liquidus temperature or for a highly heterogeneous medium, the ratios become 1:1:2.

Spherical Source

The moment tensor for a spherical source is given as

$$\mathbf{M} = (\lambda + 2\mu) \Delta V_s \begin{pmatrix} 1 & 0 & 0 \\ 0 & 1 & 0 \\ 0 & 0 & 1 \end{pmatrix}, \quad (32)$$

Figure 8a is a tensile crack with normal direction (θ, ϕ) , and we find that $v = (\sin \theta \cos \phi, \sin \theta \sin \phi, \cos \theta)$ and $[\mathbf{u}] = [u](\sin \theta \cos \phi, \sin \theta \sin \phi, \cos \theta)$. Using Eqs. (13) and (17), we obtain the moment tensor for the crack:

where $\Delta V_s = 4\pi R^2 \Delta_s$ is the volume change caused by the displacement Δ_s of a spherical surface of radius R (Müller 2001) (see “Appendix B: Moment Tensor for a Spherical Source”). The moment tensor for a spherical source may be useful to describe the source of explosion earthquakes (e.g., Kanamori et al. 1984).

Cylindrical Source

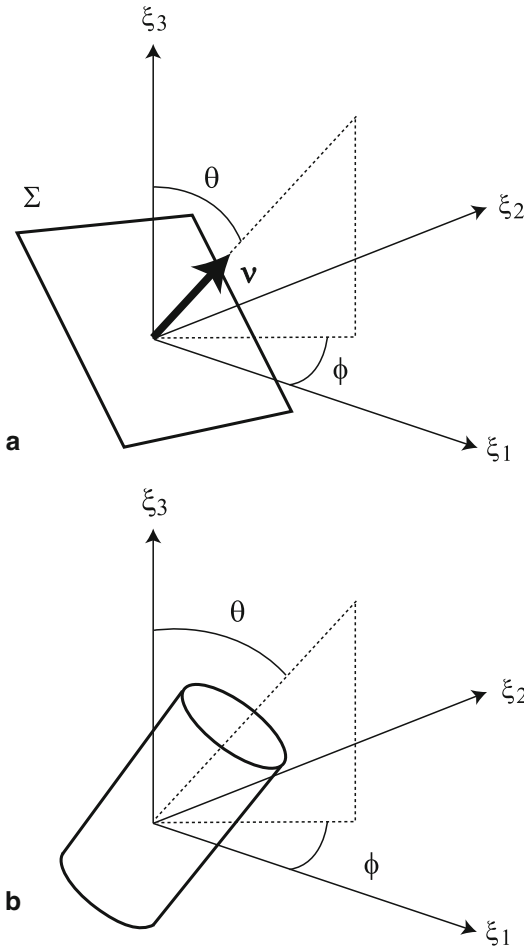
The moment tensor for a vertical cylinder is

$$\mathbf{M} = \frac{\lambda + 2\mu}{\lambda + \mu} \mu \Delta V_c \begin{pmatrix} \lambda/\mu + 1 & 0 & 0 \\ 0 & \lambda/\mu + 1 & 0 \\ 0 & 0 & \lambda/\mu \end{pmatrix}, \quad (33)$$

where $\Delta V_c = 2\pi R L \Delta_c$ is the volume change caused by the radial displacement Δ_c of the surface of a cylinder of radius R and length L (Müller 2001; Uhira and Takeo 1994) (see “Appendix C: Moment Tensor for a Cylindrical Source”).

The moment tensor for a cylinder with axis orientation angles (θ, ϕ) (Fig. 8b) is given as

$$\begin{aligned}
 \mathbf{M} &= \frac{\lambda + 2\mu}{\lambda + \mu} \mu \Delta V_c \\
 &\times \begin{pmatrix} \lambda/\mu + (\cos^2 \theta \cos^2 \phi + \sin^2 \phi) & \sin^2 \theta \sin \phi \cos \phi & \sin \theta \cos \theta \cos \phi \\ \sin^2 \theta \sin \phi \cos \phi & \lambda/\mu + (\cos^2 \theta \sin^2 \phi + \cos^2 \phi) & \sin \theta \cos \theta \sin \phi \\ \sin \theta \cos \theta \cos \phi & \sin \theta \cos \theta \sin \phi & \lambda/\mu + \sin^2 \theta \end{pmatrix} \quad (34)
 \end{aligned}$$



Source Quantification of Volcanic-Seismic Signals, Fig. 8 Source coordinates and geometries for (a) a crack and (b) a cylinder

(see “[Appendix C: Moment Tensor for a Cylindrical Source](#)”). Radial volumetric changes of a pipe may occur in association with mass transport in a cylindrical conduit (e.g., Uhira and Takeo 1994).

Inertial Glut

In the scheme presented by BM, no single force exists in indigenous seismic sources because of the requirement for conservation of linear momentum. However, Takei and Kumazawa (1994) presented a theoretical justification for the existence of a single force in indigenous sources, if the nonlinear effect of mass advection is taken into account.

Following the theory of Takei and Kumazawa (1994), we consider the exact equation of motion of the earth, given as

$$\rho^{\text{true}} \frac{D^2 u_i}{Dt^2} = \frac{\partial \sigma_{ni}^{\text{true}}}{\partial x_n}, \quad (35)$$

where ρ^{true} is true density as a function of both space and time. $\frac{D}{Dt}$ is the particle derivative:

$$\frac{D}{Dt} = \frac{\partial}{\partial t} + v_k \frac{\partial}{\partial x_k}, \quad (36)$$

where $v_k = \frac{Du_k}{Dt}$ is particle velocity. Using model density ρ^{model} and model stress $\sigma_{ni}^{\text{model}}$ from Eq. (2), Eq. (35) can be rewritten as follows:

$$\rho^{\text{model}} \frac{\partial^2 u_i}{\partial t^2} = \frac{\partial \sigma_{ni}^{\text{model}}}{\partial x_n} + f_i^V. \quad (37)$$

This equation is similar to Eq. (3), but the equivalent body force f_i^V is defined in this case as

$$\begin{aligned} f_i^V = & \left(\frac{\partial \sigma_{ni}^{\text{true}}}{\partial x_n} - \frac{\partial \sigma_{ni}^{\text{model}}}{\partial x_n} \right) \\ & + \left(\rho^{\text{model}} \frac{\partial^2 u_i}{\partial t^2} - \rho^{\text{true}} \frac{D^2 u_i}{Dt^2} \right) = f_i^S + f_i^I. \end{aligned} \quad (38)$$

Here, f_i^V is f_i^S , arising from the stress glut, plus f_i^I , representing the difference between the inertia terms, which is called the inertial glut (Takei and Kumazawa 1994).

In the same way as presented by BM, we expand f_i^I in terms of polynomial moments at a particular point ξ_0 as

$$f_i^I(\eta, \tau) = F_i^{I0}(\xi_0, \tau) \delta(\eta - \xi_0) + \text{higher order terms}, \quad (39)$$

where

$$\begin{aligned} F_i^{I0} = & \iiint_V f_i^I(\eta, \tau) dV(\eta) \\ = & - \iiint_V \left(\rho^{\text{true}} \frac{D^2 u_i}{Dt^2} - \rho^{\text{model}} \frac{\partial^2 u_i}{\partial t^2} \right) dV(\eta). \end{aligned} \quad (40)$$

By using mass conservation

$$\frac{\partial \rho^{\text{true}}}{\partial t} + \frac{\partial}{\partial x_k} (\rho^{\text{true}} v_k) = 0 \quad (41)$$

and Gauss' theorem

$$\iiint_V \frac{\partial v_k}{\partial x_k} dV = \iint_{S_e} v_k v_k dS_e, \quad (42)$$

the first integral of Eq. (40) is expressed as

$$\begin{aligned} \iiint_V \rho^{\text{true}} \frac{D^2 u_i}{Dt^2} dV &= \frac{\partial}{\partial t} \iiint_V \rho^{\text{true}} v_i dV \\ &+ \iint_{S_e} (\rho^{\text{true}} v_i v_k) v_k dS_e, \end{aligned} \quad (43)$$

where S_e is the surface of the earth. The second term in the right-hand side of Eq. (43) vanishes because it is a surface integral of a nonlinear term, which is infinitesimal. Therefore, we find that

$$\begin{aligned} \iiint_V \rho^{\text{true}} \frac{D^2 u_i}{Dt^2} dV &= \frac{\partial}{\partial t} \iiint_V \rho^{\text{true}} v_i dV \\ &= \dot{L}_i^t. \end{aligned} \quad (44)$$

Here, $\dot{L}_i^t$ is the time derivative of the total linear momentum, which is zero at any given time because of conservation of linear momentum of the earth. The second integral in Eq. (40) is given as

$$\begin{aligned} \iiint_V \rho^{\text{model}} \frac{\partial^2 u_i}{\partial t^2} dV &= \frac{\partial}{\partial t} \iiint_V \rho^{\text{model}} \frac{\partial u_i}{\partial t} dV \\ &= \dot{L}_i^m, \end{aligned} \quad (45)$$

which is the time derivative of linear momentum defined by the model density. Therefore, Eq. (40) is written as

$$F_i^{l0} = -(\dot{L}_i^t - \dot{L}_i^m) = \dot{L}_i^m, \quad (46)$$

which shows that force F_i^{l0} exists as an apparent change of the total linear momentum of the earth referred to a biased model density structure.

Takei and Kumazawa (1994) stated that F_i^{l0} originates from (a) the difference of density structure of the prescribed model from the actual value in the source region before the event and from (b) a temporal change of the density structure in the source region caused by finite displacement of

mass during the event. The latter is a nonlinear effect caused by mass advection, which may occur by fluid flow, especially in volcanic regions.

Single Force Defined by Action and Reaction

We further consider the single force from the viewpoint of the interaction force between the source volume and the rest of the earth. Following Takei and Kumazawa (1995), we define the source region V_s enclosed by a surface S and the remaining part of the earth $\bar{V}$ (Fig. 9). Using Eqs. (38), (9), (44), and (45), we obtain

$$\begin{aligned} \iiint_V f_i^V dV &= \iiint_V (f_i^S + f_i^I) dV \\ &= \iiint_V f_i^I dV \\ &= -\frac{\partial}{\partial t} \iiint_V \left(\rho^{\text{true}} v_i - \rho^{\text{model}} \frac{\partial u_i}{\partial t} \right) dV \\ &= -\frac{\partial}{\partial t} \left[\iiint_{V_s} \left(\rho^{\text{true}} v_i - \rho^{\text{model}} \frac{\partial u_i}{\partial t} \right) dV \right. \\ &\quad \left. - \iiint_{\bar{V}} \left(\rho^{\text{true}} v_i - \rho^{\text{model}} \frac{\partial u_i}{\partial t} \right) dV \right], \end{aligned} \quad (47)$$

where infinitesimal surface integral terms are omitted. The second integral in the right-hand side of Eq. (47) vanishes because $\rho^{\text{true}} = \rho^{\text{model}}$ and $v_i = \partial u_i / \partial t$ in $\bar{V}$. f_i^V exists only in the source region, and thus the volume integral of f_i^V in the left-hand side of Eq. (47) can be limited to V_s . Therefore, Eq. (47) is written as

$$\iiint_{V_s} f_i^V dV = -\frac{\partial}{\partial t} \iiint_{V_s} \left(\rho^{\text{true}} v_i - \rho^{\text{model}} \frac{\partial u_i}{\partial t} \right) dV. \quad (48)$$

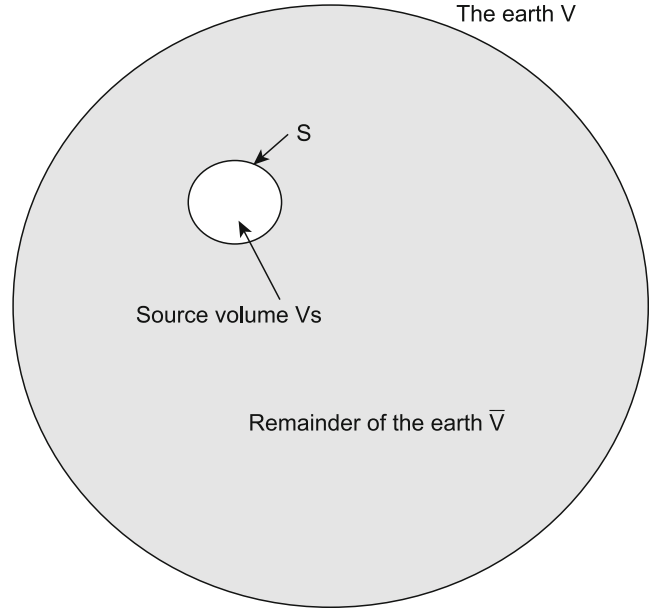
This equation is then rewritten as

$$\begin{aligned} \frac{\partial}{\partial t} \iiint_{V_s} \rho^{\text{true}} v_i dV &= \frac{\partial}{\partial t} \iiint_{V_s} \rho^{\text{model}} \frac{\partial u_i}{\partial t} dV - \iiint_{V_s} f_i^V dV \\ &= \iiint_{V_s} \left(\rho^{\text{model}} \frac{\partial^2 u_i}{\partial t^2} - f_i^V \right) dV. \end{aligned} \quad (49)$$

From Eq. (37), we find that the integrand of the right-hand side of Eq. (49) equals $\partial \sigma_{ni}^{\text{model}} / \partial x_n$. Therefore, we obtain

Source Quantification of Volcanic-Seismic Signals,

Fig. 9 The earth V divided into a source volume V_s enclosed by a surface S and the remainder of the earth $\bar{V}$



$$\begin{aligned} \frac{\partial}{\partial t} \iiint_{V_s} \rho^{\text{true}} v_i dV \\ = \iiint_{V_s} \frac{\partial \sigma_{ni}^{\text{model}}}{\partial x_n} dV \\ = \iint_S \sigma_{ni}^{\text{model}} v_n dS. \end{aligned} \quad (50)$$

This indicates that a temporal change of linear momentum in the source region is equivalent to the total force exerted on the boundary S . Therefore, $\iint_S \sigma_{ni}^{\text{model}} v_n dS$ can be regarded as the action force originating from nonlinear processes in the source volume V_s . In $\bar{V}$, the reaction force opposite to $\iint_S \sigma_{ni}^{\text{model}} v_n dS$ acts through the boundary S . We define traction P_i on S , which satisfies the following condition:

$$\sigma_{ni}^{\text{model}} v_n = -P_i \quad \text{on } S. \quad (51)$$

Then, we find that

$$\begin{aligned} \iint_S P_i dS &= - \iint_S \sigma_{ni}^{\text{model}} v_n dS \\ &= - \frac{\partial}{\partial t} \iiint_{V_s} \rho^{\text{true}} v_i dV, \end{aligned} \quad (52)$$

which is the reaction force opposite to the action force.

If the source is small compared to the wavelengths of seismic waves, the action force originating in a closed small source volume V_s may not be observable outside the source volume $\bar{V}$. Instead, the traction P_j exerted on S may be viewed in $\bar{V}$ as a traction discontinuity on S . As shown by Aki and Richards (2002), a traction discontinuity is the source of seismic waves represented by an equivalent body force. We consider seismic waves in $\bar{V}$, and the equation of motion in $\bar{V}$ may be expressed as

$$\rho^{\text{model}} \frac{\partial^2 u_i}{\partial t^2} = \frac{\partial \sigma_{ni}^{\text{model}}}{\partial x_n} + f_i \quad \text{in } \bar{V}, \quad (53)$$

where f_i is the equivalent body force due to the traction discontinuity on S , given as

$$f_i(\boldsymbol{\eta}, \tau) = \iint_S P_i(\boldsymbol{\zeta}, \tau) \delta(\boldsymbol{\eta} - \boldsymbol{\zeta}) dS(\boldsymbol{\zeta}). \quad (54)$$

Note that

$$\iiint_{\bar{V}} f_i(\boldsymbol{\eta}, \tau) dV(\boldsymbol{\eta}) = \iint_S P_i(\boldsymbol{\zeta}, \tau) dS(\boldsymbol{\zeta}), \quad (55)$$

which is the total force acting through S in $\bar{V}$.

We may use Green's functions estimated from the model structure because the effect of the true

structure in the source volume is negligible in the point source approximation. Therefore, by using Eq. (54), the displacement field u_i^f due to the single force may be described as

$$\begin{aligned} u_i^f(\mathbf{x}, t) &= \int_{-\infty}^{\infty} \iiint_{\bar{V}} f_j(\boldsymbol{\eta}, \tau) G_{ij}(\mathbf{x}, t - \tau; \boldsymbol{\eta}, 0) dV(\boldsymbol{\eta}) \tau \\ &= \int_{-\infty}^{\infty} \iiint_{\bar{V}} \left[\iiint_S P_j(\boldsymbol{\xi}, \tau) \delta(\boldsymbol{\eta} - \boldsymbol{\xi}) dS(\boldsymbol{\xi}) \right] \\ &\quad \cdot G_{ij}(\mathbf{x}, t - \tau; \boldsymbol{\eta}, 0) dV(\boldsymbol{\eta}) d\tau \\ &= \int_{-\infty}^{\infty} \iiint_S P_j(\boldsymbol{\xi}, \tau) G_{ij}(\mathbf{x}, t - \tau; \boldsymbol{\xi}, 0) dS(\boldsymbol{\xi}) d\tau, \end{aligned} \quad (56)$$

or, equivalently,

$$u_i^f(\mathbf{x}, t) = \int_{-\infty}^{\infty} F_j(\boldsymbol{\xi}_0, \tau) G_{ij}(\mathbf{x}, t - \tau; \boldsymbol{\xi}_0, 0) d\tau, \quad (57)$$

where

$$\begin{aligned} F_j(\boldsymbol{\xi}_0, \tau) &= \iint_S P_j(\boldsymbol{\xi}, \tau) dS(\boldsymbol{\xi}) \\ &= -\frac{\partial}{\partial t} \iiint_{V_s} \rho^{\text{true}} v_i dV \\ &= -\dot{L}_j^s : . \end{aligned} \quad (58)$$

Here, L_j^s is the total linear momentum inside the source volume. This single force representation in terms of the traction P_j is consistent with the relationship between the traction modes and equivalent body force given by Takei and Kumazawa (1995) in small-source and long-wavelength approximations.

F_j is the reaction force opposite to the action force generated by nonlinear processes in the source volume. A simple example is dense materials falling in a magma chamber (Fig. 10a). During the acceleration stage of falling dense materials, $\dot{L}_j^s$ is a downward force and we therefore observe an upward single force as the reaction force; during the deceleration stage, we observe a downward single force (Fig. 10a). In the case of lighter

materials ascending in a magma chamber, we observe a downward single force during the acceleration stage of ascending lighter materials, and an upward single force is observed during the deceleration stage of ascending lighter materials (Fig. 10b).

Summary

Here, we summarize the phenomenological representation of indigenous seismic sources. For point sources, the displacement field u_i can be expressed by the superposition of the displacement fields excited by single force u_i^f of Eq. (57) and moment tensor u_i^{s1} of Eq. (16):

$$\begin{aligned} u_i(\mathbf{x}, t) &= \int_{-\infty}^{\infty} F_j(\boldsymbol{\xi}_0, \tau) G_{ij}(\mathbf{x}, t - \tau; \boldsymbol{\xi}_0, 0) \tau \\ &\quad + \int_{-\infty}^{\infty} M_{jk}(\boldsymbol{\xi}_0, \tau) \frac{\partial}{\partial \xi_k} G_{ij}(\mathbf{x}, t - \tau; \boldsymbol{\xi}_0, 0) d\tau, \end{aligned} \quad (59)$$

where

$$F_j(\boldsymbol{\xi}_0, \tau) = \iint_S P_j(\boldsymbol{\xi}, \tau) dS(\boldsymbol{\xi}) \quad (60)$$

and

$$M_{jk}(\boldsymbol{\xi}_0, \tau) = \iint_{\Sigma} m_{jk}(\boldsymbol{\xi}, \tau) d\Sigma(\boldsymbol{\xi}). \quad (61)$$

Here, P_j is the traction exerted on the boundary S between the source volume and the rest of the earth, and m_{jk} is the moment density tensor defined by the displacement discontinuity across the surface Σ . F_j is related to the total linear momentum within the source volume L_j^s as

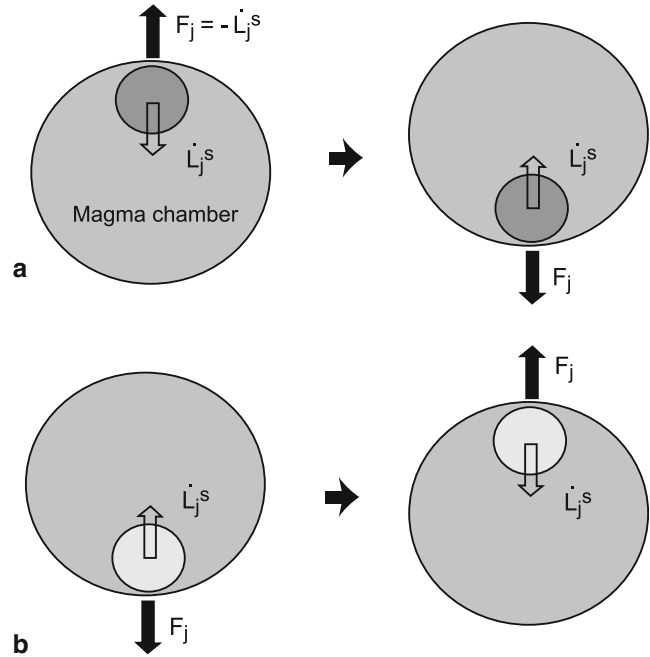
$$F_j = -\dot{L}_j^s \quad (j = 1, 2, 3). \quad (62)$$

M_{jk} is the moment tensor consisting of six independent components:

$$\mathbf{M} = \begin{pmatrix} M_{11} & M_{12} & M_{13} \\ M_{12} & M_{22} & M_{23} \\ M_{13} & M_{23} & M_{33} \end{pmatrix}. \quad (63)$$

Source Quantification of Volcanic-Seismic Signals,

Fig. 10 Schematic illustration of the reaction force (F_j) opposite to the action force ($\dot{L}_j^s$) generated by nonlinear processes in a source volume: (a) dense materials falling in a magma chamber and (b) light materials ascending in a magma chamber



Waveform Inversion

Waveform inversion is a quantitative tool used to estimate the source mechanisms and locations of volcano-seismic signals. The source parameters are determined by finding the best fits between observed and synthetic seismic waveforms. This can be performed by solving an inverse problem by using the concept of linear inverse theory (e.g., Menke 1989).

We may reasonably assume that the source of a tectonic earthquake is represented by the moment tensor corresponding to a double couple. However, volcano-seismic signals may be generated by complex source processes, requiring a general source representation consisting of single force and moment tensor. Furthermore, while we can assume a steplike function for the source-time function of a tectonic earthquake, we must determine the complex time history at the source of volcano-seismic signals, which involves more unknowns. Waveform inversion of volcano-seismic signals is thus a more complicated task than the inversion of a tectonic earthquake.

We assume a single force and moment tensor at a particular source location and obtain from Eq. (59) the following relationship:

$$u_i(t) = \int_{-\infty}^{\infty} [F_j(\tau)G_{ij}(t-\tau) + M_{jk}(\tau)G_{ij,k}(t-\tau)] d\tau, \quad (64)$$

where $G_{ij,k} = \partial G_{ij}/\partial \xi_k$. We rewrite Eq. (64) in the following form:

$$u_i(t) = \sum_{j=1}^9 \int_{-\infty}^{\infty} m_j(\tau)\Gamma_{ij}(t-\tau) d\tau, \quad (65)$$

where

$$[m_1, m_2, m_3, m_4, m_5, m_6, m_7, m_8, m_9] = [F_1, F_2, F_3, M_{11}, M_{22}, M_{33}, M_{12}, M_{23}, M_{13}] \quad (66)$$

and

$$[\Gamma_{i1}, \Gamma_{i2}, \Gamma_{i3}, \Gamma_{i4}, \Gamma_{i5}, \Gamma_{i6}, \Gamma_{i7}, \Gamma_{i8}, \Gamma_{i9}] = [G_{i1}, G_{i2}, G_{i3}, G_{i1,1}, G_{i2,2}, G_{i3,3}, G_{i1,2} + G_{i2,1}, G_{i2,3} + G_{i3,2}, G_{i1,3} + G_{i3,1}]. \quad (67)$$

We discretize t and τ as $t = k\Delta t$ ($k = 0, \dots, N_t - 1$) and $\tau = l\Delta t$ ($l = 0, \dots, N_t - 1$),

respectively, where Δt is the sampling interval. This yields the discrete form of Eq. (65) given as

$$u_i(k\Delta t) = \sum_{j=1}^9 \sum_{l=0}^{N_t-1} m_j(l\Delta t) \Gamma_{ij}(k\Delta t - l\Delta t) \Delta t : . \quad (68)$$

If we have N_c traces of u_i , Eq. (68) is written in matrix form

$$\mathbf{u} = \Gamma \mathbf{m} : , \quad (69)$$

where

$$\mathbf{u} = [u_1(0), \dots, u_1((N_t - 1)\Delta t), \dots, u_{N_c}(0), \dots, u_{N_c}((N_t - 1)\Delta t)]^T, \quad (70)$$

$$\mathbf{m} = [m_1(0), \dots, m_1((N_t - 1)\Delta t), \dots, m_9(0), \dots, m_9((N_t - 1)\Delta t)]^T, \quad (71)$$

$$\Gamma = \begin{bmatrix} \Gamma_{11}(0) & \cdots & \Gamma_{11}((1 - N_t)\Delta t) & \cdots & \cdots & \Gamma_{19}(0) & \cdots & \Gamma_{19}((1 - N_t)\Delta t) \\ \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots \\ \Gamma_{11}((N_t - 1)\Delta t) & \cdots & \Gamma_{11}(0) & \cdots & \cdots & \Gamma_{19}((N_t - 1)\Delta t) & \cdots & \Gamma_{19}(0) \\ \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots \\ \Gamma_{N_c1}(0) & \cdots & \Gamma_{N_c1}((1 - N_t)\Delta t) & \cdots & \cdots & \Gamma_{N_c9}(0) & \cdots & \Gamma_{N_c9}((1 - N_t)\Delta t) \\ \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots & \vdots \\ \Gamma_{N_c1}((N_t - 1)\Delta t) & \cdots & \Gamma_{N_c1}(0) & \cdots & \cdots & \Gamma_{N_c9}((N_t - 1)\Delta t) & \cdots & \Gamma_{N_c9}(0) \end{bmatrix} \Delta t. \quad (72)$$

Here T denotes transpose. Γ is called the data kernel, which is an $N \times M$ matrix, and $\mathbf{u}$ and $\mathbf{m}$ are vectors of lengths N and M , respectively, where $N = N_c N_t$ and $M = 9 N_t$.

We denote observed waveforms as $\mathbf{u}^{\text{obs}}$, which satisfy the following equation:

$$\mathbf{u}^{\text{obs}} - \Gamma \mathbf{m} = \mathbf{e}, \quad (73)$$

where $\mathbf{e}$ is the prediction error or misfit. In the case of $N > M$, we estimate $\mathbf{m}$ by minimizing the residual R , defined as

$$R = \frac{\mathbf{e}^T \mathbf{e}}{\mathbf{u}^{\text{obs}T} \mathbf{u}^{\text{obs}}} = \frac{\sum_{i=1}^N e_i^2}{\sum_{i=1}^N (u_i^{\text{obs}})^2}, \quad (74)$$

where

$$\mathbf{e}^T \mathbf{e} = (\mathbf{u}^{\text{obs}} - \Gamma \mathbf{m})^T (\mathbf{u}^{\text{obs}} - \Gamma \mathbf{m}). \quad (75)$$

To minimize R is to find values of $\mathbf{m}$ that satisfy the following equations:

$$\frac{\partial R}{\partial m_q} = 0 \quad (q = 1, \dots, M). \quad (76)$$

This leads to the following relationship (the normal equation):

$$\Gamma^T \Gamma \mathbf{m} - \Gamma^T \mathbf{u}^{\text{obs}} = 0. \quad (77)$$

Therefore, we have the following solution:

$$\mathbf{m}^{\text{est}} = [\Gamma^T \Gamma]^{-1} \Gamma^T \mathbf{u}^{\text{obs}}, \quad (78)$$

which is the least-squares solution. This least-squares approach in the time domain has been used in many volcano seismological studies (e.g., Chouet et al. 2003, 2005; Hidayat et al. 2002; Kumagai et al. 2001, 2002b, 2003, 2005; Nakano and Kumagai 2005b; Nakano et al. 2003; Nishimura et al. 2000; Ohminato 2006; Ohminato et al. 1998, 2006; Takeo et al. 1990; Tameguri et al. 2002).

As can be seen from Eq. (68), u_i is calculated by the convolution of m_j with Γ_{ij} . This means that the displacement $u_i(k\Delta t)$ includes the entire effect of the time history of the single force and moment

tensor as well as Green's functions before time $k\Delta t$. This may be easily understood if we consider $u_i((N_t - 1)\Delta t)$ in Eq. (68):

$$\begin{aligned} u_i((N_t - 1)\Delta t) &= \Delta t \sum_j \\ &= 1^9 [m_j(0)\Gamma_{ij}((N_t - 1)\Delta t) \\ &\quad + m_j(\Delta t)\Gamma_{ij}((N_t - 2)\Delta t) + \dots \\ &\quad + m_j((N_t - 2)\Delta t)\Gamma_{ij}(\Delta t) \\ &\quad + m_j((N_t - 1)\Delta t)\Gamma_{ij}(0)]. \end{aligned} \quad (79)$$

The convolution relationship leads to the large matrix of Γ of Eq. (69), which requires long computation time to obtain the least-squares solution. A better alternative is to solve the inverse problem in the frequency domain (Auger et al. 2006; Nakano et al. 2007; Stump and Johnson 1977). The Fourier transform of Eq. (65) is given by

$$\bar{u}_i(\omega) = \sum_{j=1}^9 \bar{m}_j(\omega) \bar{\Gamma}_{ij}(\omega), \quad (80)$$

where ω is the angular frequency and $\bar{u}_i, \bar{m}_j$, and $\bar{\Gamma}_{ij}$ are the Fourier transforms of u_i, m_j , and Γ_{ij} . This frequency domain equation is discretized as follows:

$$\bar{u}_i(k\Delta\omega) = \sum_{j=1}^9 \bar{m}_j(k\Delta\omega) \bar{\Gamma}_{ij}(k\Delta\omega), \quad (81)$$

where $\Delta\omega = 2\pi/(N_t\Delta t)$ and $k = 0, \dots, N_t/2$. Note that the Fourier-transform components are complex and we have a set of the equations corresponding to real and imaginary parts; therefore, Eq. (81) consists of N_t equations. We further note that $\bar{u}_i, \bar{m}_j$, and $\bar{\Gamma}_{ij}$ in Eq. (81) depend only on $k\Delta\omega$. This indicates that individual frequency components of $\bar{u}_i, \bar{m}_j$, and $\bar{\Gamma}_{ij}$ are independent of each other and Eq. (81) can be solved separately. This is different from the time-domain Eq. (65): u_i is a function of $k\Delta t$, whereas m_j is a function of $l\Delta t$, and Γ_{ij} is a function of $k\Delta t$ and $l\Delta t$, so we cannot separate the time-domain equation in individual values of $k\Delta t$. The use of the frequency-domain equation results in a smaller size of the data kernel, as shown below.

If we have N_c traces of $\bar{u}_i$, Eq. (81) is written in matrix form:

$$\bar{\mathbf{u}} = \bar{\mathbf{\Gamma}} \bar{\mathbf{m}}. \quad (82)$$

Here, $\bar{\mathbf{u}}$ is a vector consisting of $\bar{u}_i$, which is arranged in individual frequency components as follows:

$$\begin{aligned} \bar{\mathbf{u}} &= [\bar{u}_1(0), \bar{u}_2(0), \dots, \bar{u}_{N_c}(0), \bar{u}_1(\Delta\omega), \\ &\quad \bar{u}_2(\Delta\omega), \dots, \bar{u}_{N_c}(\Delta\omega), \dots, \bar{u}_1((N_t/2)\Delta\omega), \\ &\quad \bar{u}_2((N_t/2)\Delta\omega), \dots, \bar{u}_{N_c}((N_t/2)\Delta\omega)]^T \end{aligned} \quad (83)$$

or, equivalently,

$$\bar{\mathbf{u}} = [\bar{\mathbf{u}}(0), \bar{\mathbf{u}}(\Delta\omega), \dots, \bar{\mathbf{u}}(k\Delta\omega), \dots, \bar{\mathbf{u}}(N_t/2\Delta\omega)]^T, \quad (84)$$

where

$$\bar{\mathbf{u}}(k\Delta\omega) = [\bar{u}_1(k\Delta\omega), \bar{u}_2(k\Delta\omega), \dots, \bar{u}_{N_c}(k\Delta\omega)]^T. \quad (85)$$

Similarly, $\bar{\mathbf{m}}$ is a vector consisting of $\bar{m}_j$, which is arranged in individual frequency components as

$$\begin{aligned} \bar{\mathbf{m}} &= [\bar{m}_1(0), \bar{m}_2(0), \dots, \bar{m}_9(0), \bar{m}_1(\Delta\omega), \bar{m}_2(\Delta\omega), \dots, \\ &\quad \bar{m}_9(\Delta\omega), \dots, \bar{m}_1((N_t/2)\Delta\omega), \bar{m}_2((N_t/2)\Delta\omega), \dots, \\ &\quad \bar{m}_9((N_t/2)\Delta\omega)]^T \end{aligned} \quad (86)$$

or

$$\bar{\mathbf{m}} = [\bar{\mathbf{m}}(0), \bar{\mathbf{m}}(\Delta\omega), \dots, \bar{\mathbf{m}}(k\Delta\omega), \dots, \bar{\mathbf{m}}((N_t/2)\Delta\omega)]^T, \quad (87)$$

where

$$\bar{\mathbf{m}}(k\Delta\omega) = [\bar{m}_1(k\Delta\omega), \bar{m}_2(k\Delta\omega), \dots, \bar{m}_9(k\Delta\omega)]^T. \quad (88)$$

Accordingly, $\bar{\mathbf{\Gamma}}$ is given as

$$\bar{\mathbf{\Gamma}} = \begin{bmatrix} \bar{\Gamma}(0) & & & & \\ & \bar{\Gamma}(\Delta\omega) & & & \\ & & \ddots & & \\ & & & \bar{\Gamma}(k\Delta\omega) & \\ & 0 & & & \ddots \\ & & & & & \bar{\Gamma}((N_t/2)\Delta\omega) \end{bmatrix}, \quad (89)$$

where $\bar{\Gamma}(k\Delta\omega)$ is

$$\bar{\Gamma}(k\Delta\omega) = \begin{bmatrix} \bar{\Gamma}_{11}(k\Delta\omega) & \cdots & \bar{\Gamma}_{19}(k\Delta\omega) \\ \vdots & \ddots & \vdots \\ \bar{\Gamma}_{N_c,1}(k\Delta\omega) & \cdots & \bar{\Gamma}_{N_c,9}(k\Delta\omega) \end{bmatrix}. \quad (90)$$

Since $\bar{\Gamma}$ is the block diagonal matrix, Eq. (82) can be reduced to

$$\bar{\mathbf{u}}(k\Delta\omega) = \bar{\Gamma}(k\Delta\omega) \bar{\mathbf{m}}(k\Delta\omega) \quad (k = 0, \dots, N_t/2). \quad (91)$$

The dimension of matrix $\bar{\Gamma}(k\Delta\omega)$ is $N_c \times 9$, which is N_t^2 times smaller than that of Γ in the time domain (Eq. (72)). This is because convolution in the time domain is equivalent to the multiplication in the frequency domain. The smaller size of $\bar{\Gamma}(k\Delta\omega)$ contributes to faster computation of the least-squares solution. We denote the $k\Delta\omega$ component of the Fourier transform of observed seismograms as

$$\bar{\mathbf{u}}^{\text{obs}}(k\Delta\omega) = [\bar{u}_1^{\text{obs}}(k\Delta\omega), \bar{u}_2^{\text{obs}}(k\Delta\omega), \dots, \bar{u}_{N_c}^{\text{obs}}(k\Delta\omega)]^T. \quad (92)$$

The least-squares solution in the frequency domain for the $k\Delta\omega$ component is then given by

$$\bar{\mathbf{m}}^{\text{est}}(k\Delta\omega) = [\bar{\Gamma}^H(k\Delta\omega) \bar{\Gamma}(k\Delta\omega)]^{-1} \bar{\Gamma}^H(k\Delta\omega) \bar{\mathbf{u}}^{\text{obs}}(k\Delta\omega), \quad (93)$$

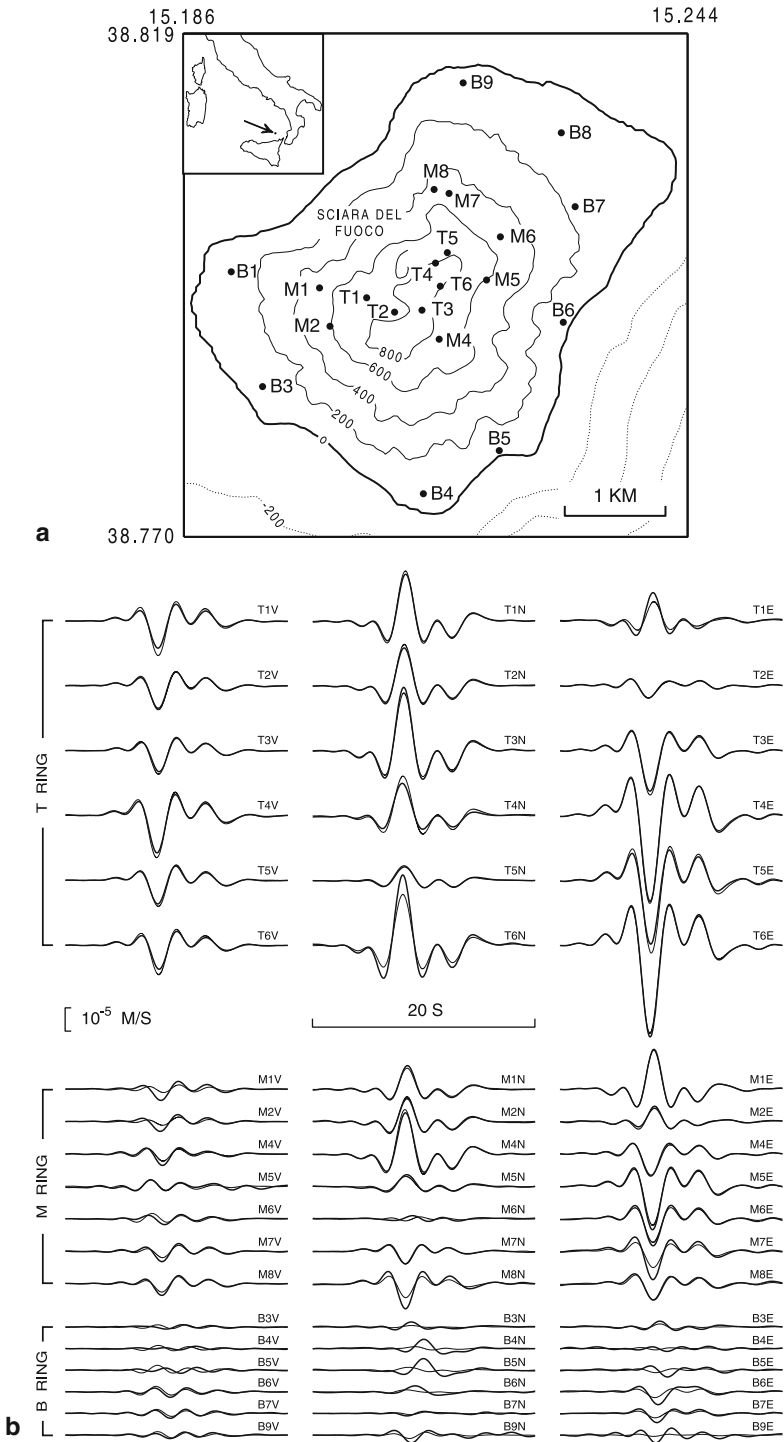
where H denotes the adjoint (Hermitian conjugate). We perform the inverse Fourier transform of $\bar{\mathbf{m}}^{\text{est}}(k\Delta\omega)$ ($k = 0, \dots, N_t/2$) to obtain the solution in the time domain.

The use of appropriate Green's functions is critically important to obtain the correct solution $\mathbf{m}^{\text{est}}$ in either the time or frequency domains. If the source is located in a medium that can be approximated by a homogeneous or layered structure with a flat surface, Green's functions can be calculated by using propagator matrices (e.g., Kennet and Kerry 1979) and the discrete wavenumber method (e.g., Bouchon 1979, 1981; Chouet 1982). However, this situation is not common in volcanic regions, and we need to take into account the steep topographies and heterogeneous structures of volcanoes.

The boundary integral equation method (e.g., Bouchon et al. 1996) and boundary element method (e.g., Kawase 1988) can be used to quantify the effect of three-dimensional topography in a homogeneous or layered structure. To treat both topography and structural heterogeneity simultaneously, the finite-difference method (e.g., Ohminato and Chouet 1997; Ripperger et al. 2003) and the discrete lattice method (O'Brien and Bean 2004) are suitable. Once Green's functions (and their spatial derivatives) are calculated, we can obtain the least-squares solution $\mathbf{m}^{\text{est}}$ through Eq. (78) or (93). This solution is, however, obtained for a particular source point, and a grid search in space is required to find the best-fit source location.

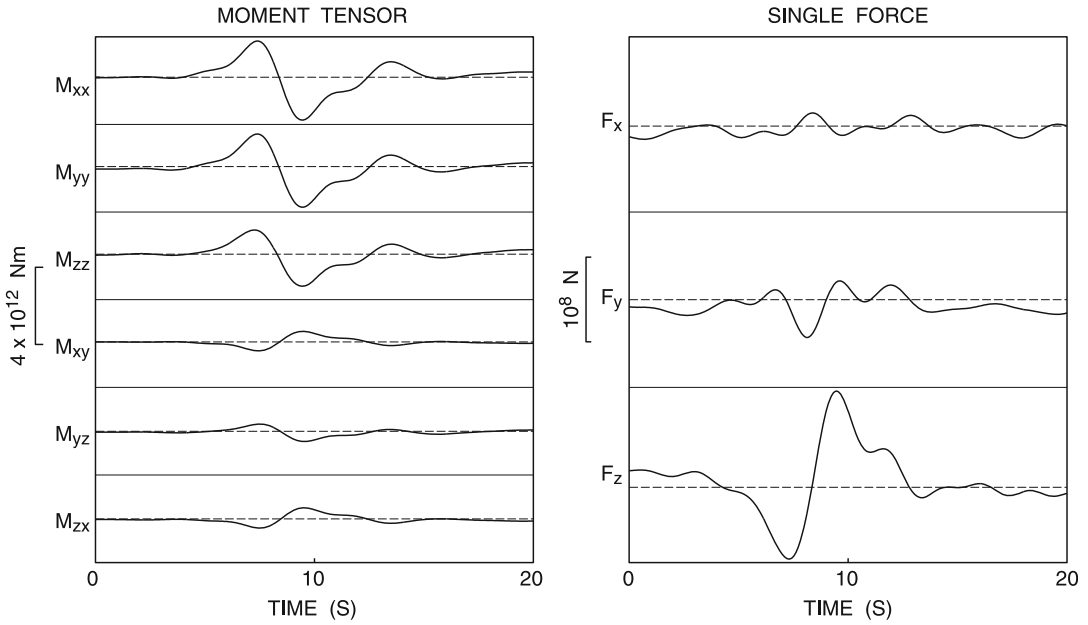
An example of waveform inversion is provided in Figs. 11 and 12. The source mechanism of VLP signals observed by a dense broadband seismic network at Stromboli volcano, Italy, was estimated by using the waveform inversion method in the time domain (Chouet et al. 2003). In the inversion, VLP waveforms from 18 stations (Fig. 11) were used, and 6 moment tensors and 3 single force components for a point source were assumed. Green's functions were calculated by the finite-difference method (Ohminato and Chouet 1997) using the topography of the volcano. The inversion result (Fig. 12) points to volumetric changes accompanying a dominantly vertical single force at the source of the VLP signals. The mechanism is interpreted as the movement of a magma column in response to the pistonlike rise of a slug of gas in an inclined crack-like conduit of Stromboli volcano (Chouet et al. 2003).

The waveform inversion approach discussed above might be successful if we have many (more than 10) stations covering the entire edifice of a volcano. However, such dense seismic networks exist at only a limited number of volcanoes. Many volcanoes are monitored by networks of just a few seismic stations. It is also possible that signals would be detected by only a few stations close to the source even in a dense seismic network. Nakano and Kumagai (2005b) proposed a practical approach for quantifying the source of volcano-seismic signals observed by a small number of seismic stations. In this approach, possible source geometries such as a crack and pipe are assumed as a priori information in waveform inversion. This assumption reduces the number of free parameters



Source Quantification of Volcanic-Seismic Signals, Fig. 11 (a) Locations of broadband seismic stations at Stromboli volcano, Italy. (b) Observed VLP velocity waveforms (*thick lines*) bandpassed between 2 and 20 s

and the best-fit synthetic seismograms (*thin lines*) obtained from waveform inversion assuming six moment tensors and three single force components for the source mechanism (Chouet et al. 2003)



Source Quantification of Volcanic-Seismic Signals, Fig. 12 Source-time functions of six moment tensors and three single force components estimated by waveform inversion for the VLP signals shown in Fig. 11b (Chouet et al. 2003)

and enables us to estimate the source mechanism and location with waveform data from two to three three-component seismic stations. Their approach is summarized as follows. The moment tensors for a crack and pipe are given in Eqs. (30) and (34), respectively. If we assume a value for λ/μ , free parameters in the moment tensors are the source-time function, θ , and ϕ . We can determine the source-time function by the waveform inversion discussed above, in which a grid search with respect to θ and ϕ is conducted to find the best-fit angles at a fixed location. A grid search in space is also conducted to find the combination of the best-fit angles and source location, and their residuals are compared to adopt the best model. The applicability of this simple approach has been demonstrated by synthetic tests and observed seismograms (Nakano and Kumagai 2005b).

Spectral Analysis

Volcano-seismic signals often show harmonic oscillatory signatures, which are interpreted as oscillations of fluid-filled resonators at the source of these

signals. To determine the frequency characteristics of observed oscillations is essential to quantify the characteristic properties of the resonator systems. The discrete Fourier transform (DFT) has traditionally been used for this purpose. In DFT, the discrete time series $u(k\Delta t)$ ($k = 0, \dots, N_t - 1$) is transformed into the discrete frequency series

$$\bar{u}(l\Delta\omega) = \Delta t \sum_{k=0}^{N_t-1} u(k\Delta t) e^{2\pi i l k / N_t} \quad (94)$$

$$(l = 0, \dots, N_t - 1),$$

where $\Delta\omega = 2\pi/(N_t\Delta t)$ and $i = \sqrt{-1}$. DFT is the decomposition of time series into a linear combination of the orthogonal basis functions $e^{2\pi i l k / N_t}$ in the frequency domain. Because of the stationary nature of the basis functions, DFT is especially useful to analyze stationary periodic signals. However, volcano-seismic signals often show decaying harmonic oscillations. These may be viewed as the impulsive response of a resonator system, which inherently contains dissipation mechanisms. A decaying harmonic oscillation can be represented by the complex frequency, which

cannot be directly estimated by DFT based on the real frequency. The complex frequencies can be determined by solving an inverse problem for parameters in an appropriate model of a given time series. The Sompi method (Kumazawa et al. 1990) is one such approach that is based on an autoregressive (AR) model of a linear dynamic system. This method has been successfully used to determine the complex frequencies of decaying harmonic oscillations in LP and VLP events. We follow the theory of Sompi presented by Kumazawa et al. (1990) and Hori et al. (1989).

Let us consider the equation of motion for a linear dynamic system consisting of a block and spring:

$$b \frac{d^2}{dt^2} x(t) + v \frac{d}{dt} x(t) + \kappa x(t) = 0, \quad (95)$$

where b is the mass of a block, κ is the spring constant, and $v(dx(t)/dt)$ represents a dissipation term caused by friction, which is proportional to velocity. We set $x(t) = e^{i\omega t}$ in Eq. (95), which is an eigenfunction of this system. We then obtain

$$b(i\omega)^2 + v(i\omega) + \kappa = 0. \quad (96)$$

This is the characteristic equation of the system. If we consider oscillation solution ($v^2 - 4b\kappa < 0$), Eq. (96) has two characteristic solutions:

$$i\omega_1 = \frac{-v + i\sqrt{4b\kappa - v^2}}{2b}, \quad i\omega_2 = i\omega_1^*, \quad (97)$$

where $*$ denotes the complex conjugate. If we set $\omega = 2\pi(f - ig)$, we obtain

$$f_1 = -f_2 = \frac{1}{2\pi} \sqrt{\frac{\kappa}{b} - \frac{v^2}{4b^2}}, \quad (98)$$

$$g_1 = g_2 = -\frac{1}{2\pi} \frac{v}{2b}. \quad (99)$$

Let us consider the following difference equation:

$$\sum_{j=0}^2 a(j)x((k-j)\Delta t) = 0 \quad (100)$$

or

$$\begin{aligned} & a(0)x(k\Delta t) + a(1)x(k\Delta t - \Delta t) \\ & + a(2)x(k\Delta t - 2\Delta t) \\ & = 0. \end{aligned} \quad (101)$$

If we set $x(k\Delta t) = z^k$, where $z = e^{i\omega\Delta t}$, Eq. (100) is given as

$$a(0)z^k + a(1)z^{k-1} + a(2)z^{k-2} = 0 \quad (102)$$

or, equivalently,

$$a(0)z^2 + a(1)z + a(2) = 0. \quad (103)$$

Equation (103) is the characteristic equation of the difference equation. If we consider oscillation solutions ($a^2(1) - 4a(0)a(2) < 0$), Eq. (103) has the following two characteristic solutions:

$$\begin{aligned} z_1 &= \frac{-a(1) + i\sqrt{4a(0)a(2) - a^2(1)}}{2a(0)}, \\ z_2 &= z_1^*. \end{aligned} \quad (104)$$

These two solutions constitute complex conjugate pairs. By setting $\omega = 2\pi(f - ig)$, we obtain

$$\begin{aligned} f_1 &= -f_2 \\ &= \frac{1}{2\pi\Delta t} \tan^{-1} \left(-\sqrt{\frac{4a(0)a(2)}{a^2(1)} - 1} \right), \end{aligned} \quad (105)$$

$$g_1 = g_2 = \frac{1}{2\pi\Delta t} \ln \sqrt{\frac{a(2)}{a(0)}}. \quad (106)$$

These solutions are equivalent to the solutions of the linear dynamic system given in Eqs. (98) and (99). Therefore, Eq. (100) is a difference form of the equation of motion (95).

We further consider a dynamic system consisting of two block-spring subsystems,

which are coupled to each other. The equations of motion for this system are written as

$$M_1 \frac{d^2}{dt^2} x_1(t) + L_1 \frac{d}{dt} x_1(t) + K_1 x_1(t) = C(x_2 - x_1), \quad (107)$$

$$M_2 \frac{d^2}{dt^2} x_2(t) + L_2 \frac{d}{dt} x_2(t) + K_2 x_2(t) = C(x_1 - x_2), \quad (108)$$

where M_i , L_i , and K_i are the mass, dissipation coefficient, and spring constant of the i th block and C is the spring constant between the blocks. The set of the equations of motion (107) and (108) can be equivalently given in the following form:

$$\begin{pmatrix} M_1 \frac{d^2}{dt^2} + L_1 \frac{d}{dt} + K_1 + C & -C \\ -C & M_2 \frac{d^2}{dt^2} + L_2 \frac{d}{dt} + K_2 + C \end{pmatrix} \cdot \begin{pmatrix} x_1(t) \\ x_2(t) \end{pmatrix} = 0. \quad (109)$$

These equations have nontrivial solutions of $x_i(t)$ if, and only if,

$$\det \begin{vmatrix} M_1 \frac{d^2}{dt^2} + L_1 \frac{d}{dt} + K_1 + C & -C \\ -C & M_2 \frac{d^2}{dt^2} + L_2 \frac{d}{dt} + K_2 + C \end{vmatrix} = 0. \quad (110)$$

This leads to the following fourth-order differential equation:

$$\left[\begin{pmatrix} M_1 \frac{d^2}{dt^2} + L_1 \frac{d}{dt} + K_1 + C \\ M_2 \frac{d^2}{dt^2} + L_2 \frac{d}{dt} + K_2 + C \end{pmatrix} - C^2 \right] x_i(t) = 0, \quad (111)$$

where $i = 1, 2$. Each fourth-order differential equation has four complex frequencies, of which two are complex conjugates. Therefore, each

block has two independent complex frequencies in this dynamic system. Similarly, a dynamic system consisting of m block-spring subsystems, which are all connected to each other, can be described by the following $2m$ -order differential equation:

$$\sum_{j=0}^{2m} b_j \frac{d^j}{dt^j} x_i(t) = 0 \quad (i = 1, 2, \dots, m), \quad (112)$$

where b_j denotes the coefficients determined by the characteristic properties of the system. Each block has m independent complex frequencies in this case. A difference form of Eq. (112) is

$$\sum_{j=0}^{2m} a(j) x_i(k\Delta t - j\Delta t) = 0, \quad (113)$$

which is a homogeneous $2m$ -order AR equation. Although we have considered block-spring systems as an example, Eq. (112) is a general equation to describe the free oscillations of any linear systems. We use its difference form, Eq. (113), as the basic equation to determine the characteristic complex frequencies of linear systems from time series records.

Hereafter, I set $\Delta t = 1$ to simplify the following notations. We assume that observed discrete time series $u(k)$ consists of signal $x(k)$ and white noise $e(k)$:

$$u(k) = x(k) + e(k) \quad (k = 0, 1, \dots, N_t - 1), \quad (114)$$

and $x(k)$ satisfies the following homogeneous AR equation:

$$\sum_{j=0}^{2m} a(j) x(k - j) = 0. \quad (115)$$

To determine the $2m + 1$ AR coefficients, we minimize the residual R :

$$R = \frac{1}{N_t - 2m} \sum_{k=2m}^{N_t-1} \left[\sum_{j=0}^{2m} a(j)u(k-j) \right]^2. \quad (116)$$

Since Eq. (115) is homogeneous, we cannot determine absolute values of the AR coefficients. To avoid the trivial solution $a(j) = 0$ ($j = 0, 1, \dots, 2m$), we impose the condition

$$\sum_{j=0}^{2m} a^2(j) = 1. \quad (117)$$

To minimize the residual R under the condition of Eq. (117) is a conditional minimization problem. This problem can be solved by minimizing the residual R' :

$$R' = R - \beta \left(\sum_{j=0}^{2m} a^2(j) - 1 \right), \quad (118)$$

where β is a constant. Consequently, we solve the following equation:

$$\frac{\partial}{\partial a(l)} R' = 0 \quad (l = 0, 1, \dots, 2m). \quad (119)$$

This equation leads to an eigenvalue problem

$$\mathbf{P}\mathbf{a} = \beta\mathbf{a}, \quad (120)$$

where $\mathbf{a} = (a(0), a(1), \dots, a(2m))^T$ and $\mathbf{P}$ is a non-Toeplitz matrix given by

$$\mathbf{P} = \frac{1}{N_t - 2m} \begin{bmatrix} \sum_{n=2m}^{N_t-1} u(n)u(n) & \cdots & \sum_{n=2m}^{N_t-1} u(n-2m)u(n) \\ \vdots & \ddots & \vdots \\ \sum_{n=2m}^{N_t-1} u(n)u(n-2m) & \cdots & \sum_{n=2m}^{N_t-1} u(n-2m)u(n-2m) \end{bmatrix}. \quad (121)$$

By solving the eigenvalue problem of Eq. (120), we obtain $2m + 1$ eigenvalues $\beta^{(r)}$ ($r = 1, 2, \dots, 2m + 1$) and corresponding eigenvectors $\mathbf{a}^{(r)}$ ($r = 1, 2, \dots, 2m + 1$). Note that the residual R in Eq. (116) is given as

$$R = \mathbf{a}^T \mathbf{P} \mathbf{a} = \mathbf{a}^T \beta \mathbf{a} = \beta. \quad (122)$$

Therefore, each eigenvector gives a local minimum of the residual R . We therefore adopt the eigenvector corresponding to the minimum eigenvalue, which minimizes the residual R . Hereafter, $\mathbf{a}$ denotes the eigenvector corresponding to the minimum eigenvalue.

The next step is to determine the complex frequencies from the eigenvector $\mathbf{a}$. The characteristic equation of the 2 m-order homogeneous AR equation is given as

$$\sum_{j=0}^{2m} a(j)z^{-j} = 0. \quad (123)$$

This equation is the $2m$ -degree algebraic equation for z^{-1} and has $2m$ complex solutions, which can be solved by a numerical procedure. As mentioned before, a complex solution is always accompanied by its complex conjugate, and therefore there are m independent solutions z_l ($l = 1, 2, \dots, m$) in Eq. (123). These solutions are related to the complex frequencies ($f_l - ig_l$) as

$$z_l = e^{2\pi i(f_l - ig_l)} \quad (124)$$

or

$$f_l - ig_l = \ln z_l / (2\pi i). \quad (125)$$

The Q factor is defined through the complex frequencies as

$$Q_l = f_l / (-2g_l). \quad (126)$$

The final step is to determine the amplitudes of the solution z_l . If we set a conjugate pair as $z_l = z_{m+l}^*$, the general solution of Eq. (123) is given as

$$\begin{aligned} x(k) &= \sum_{l=1}^m (\alpha_l z_l^k + \alpha_{m+l} z_{m+l}^k) \\ &= \sum_{l=1}^m (\alpha_l z_l^k + (\alpha_l z_l^k)^*). \end{aligned} \quad (127)$$

Here α_l is the complex amplitude given as

$$\alpha_l = r_l e^{i\phi_l}, \quad (128)$$

where r_l and ϕ_l are the real amplitude and phase, respectively. This is equivalently written as

$$\alpha_l = \alpha_l^r + i\alpha_l^i, \quad (129)$$

where α_l^r and α_l^i are the real and imaginary parts of α_l , respectively. Then, the general solution of Eq. (127) can be rewritten

$$x(k) = \sum_{l=1}^m (2\alpha_l^r C_l(k) - 2\alpha_l^i S_l(k)), \quad (130)$$

where $C_l(k) = e^{2\pi g_l k} \cos(2\pi f_l k)$ and $S_l(k) = e^{2\pi g_l k} \sin(2\pi f_l k)$. The observed time series thus satisfies the following equation:

$$\begin{aligned} u(k) - \sum_{l=1}^m (2\alpha_l^r C_l(k) - 2\alpha_l^i S_l(k)) &= e(k) \\ (k = 0, 1, \dots, N_t - 1) \end{aligned} \quad (131)$$

or in matrix form

$$\mathbf{u} - \mathbf{A}\boldsymbol{\alpha} = \mathbf{e}, \quad (132)$$

where

$$\mathbf{u} = (u(0), u(1), \dots, u(N_t - 1))^T, \quad (133)$$

$$\boldsymbol{\alpha} = (2\alpha_1^r, \dots, 2\alpha_m^r, 2\alpha_1^i, \dots, 2\alpha_m^i)^T, \quad (134)$$

$$\mathbf{e} = (e(0), e(1), \dots, e(N_t - 1))^T, \quad (135)$$

and

$$\mathbf{A} = \begin{bmatrix} C_1(0) & \cdots & C_m(0) & -S_1(0) & \cdots & -S_m(0) \\ \vdots & & \vdots & \vdots & & \vdots \\ C_1(N_t - 1) & \cdots & C_m(N_t - 1) & -S_1(N_t - 1) & \cdots & -S_m(N_t - 1) \end{bmatrix}. \quad (136)$$

We estimate $\boldsymbol{\alpha}$ by minimizing the residual R_a as follows:

$$R_a = (\mathbf{u} - \mathbf{A}\boldsymbol{\alpha})^T (\mathbf{u} - \mathbf{A}\boldsymbol{\alpha}). \quad (137)$$

This is an ordinary least-squares problem, and the solution is given as

$$\boldsymbol{\alpha} = [\mathbf{A}^T \mathbf{A}]^{-1} \mathbf{A}^T \mathbf{u}. \quad (138)$$

The real amplitudes r_l and phases ϕ_l at $k = 0$ are determined from estimated $\boldsymbol{\alpha}$ as

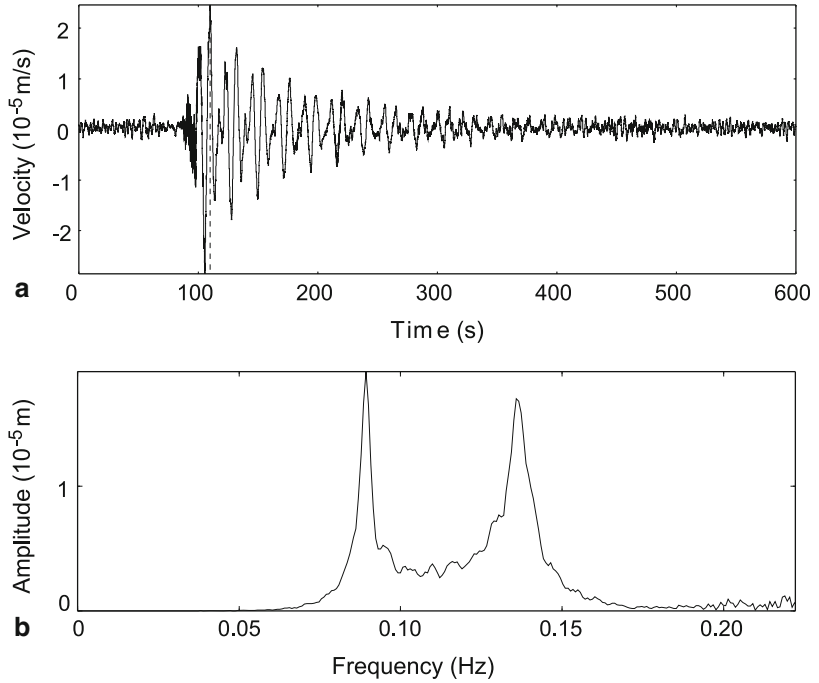
$$r_l = \sqrt{(\alpha_l^r)^2 + (\alpha_l^i)^2} \quad (l = 1, \dots, m), \quad (139)$$

$$\phi_l = \tan^{-1}(\alpha_l^i / \alpha_l^r) \quad (l = 1, \dots, m). \quad (140)$$

An example of the Sompi analysis of a VLP waveform (Fig. 13a) is shown in Fig. 14. The VLP waveform in Fig. 13a shows decaying harmonic oscillations, which are interpreted as the resonances of a magmatic dike triggered by a time-localized excitation (Kumagai 2006; Kumagai et al. 2003). The amplitude spectrum of the VLP waveform based on DFT is shown in Fig. 13b, which indicates two dominant peaks. As such, DFT provides frequency and amplitude information simultaneously. Q values may be determined by the bandwidths of these spectral peaks at halves their peak amplitudes. The estimated Q values, however, can be biased, because the

Source Quantification of Volcanic-Seismic Signals,

Fig. 13 (a) VLP waveform observed at Hachijo Island, Japan, associated with a VT earthquake swarm (Kumagai 2006; Kumagai et al. 2003). (b) Amplitude spectrum of the waveform displayed in Fig. 13a



whole waveform including the tail of decaying oscillations as well as the onset of growing oscillations is used in this approach. A box-car window can be applied to extract the decaying oscillations, but this distorts the spectrum and may lead to biased estimations of Q values.

In Sompi, on the other hand, we delete the onset portion of the signal and use the decaying harmonic oscillations only, which satisfy the homogeneous assumption of the AR equation. Since the number of signals (oscillations) is not known a priori, we use trial AR orders and solve the eigenvalue problems for individual trial AR orders to determine the complex frequencies, which are plotted in a frequency-growth rate (f-g) diagram. Figure 14a is the f-g diagram of the complex frequencies determined from the VLP waveform by using trial AR orders between 4 and 60. Densely populated regions on the f-g diagram represent the signal for which the complex frequencies are stably determined for different AR orders, while scattered points represent incoherent noise. Sompi then determines the complex amplitudes $re^{i\phi}$ by using the estimated complex frequencies. The optimum AR order and the number of signals may be determined by the two-parameter Akaike information criterion (AIC) (Matsuura et al. 1990) or the extended information criterion (EIC) (Ulrych and

Sacchi 1995). Figure 14b, c plot the amplitudes r and phases ϕ , respectively, at the beginning of the decaying harmonic oscillations and using the two complex frequencies determined at an AR order of 40.

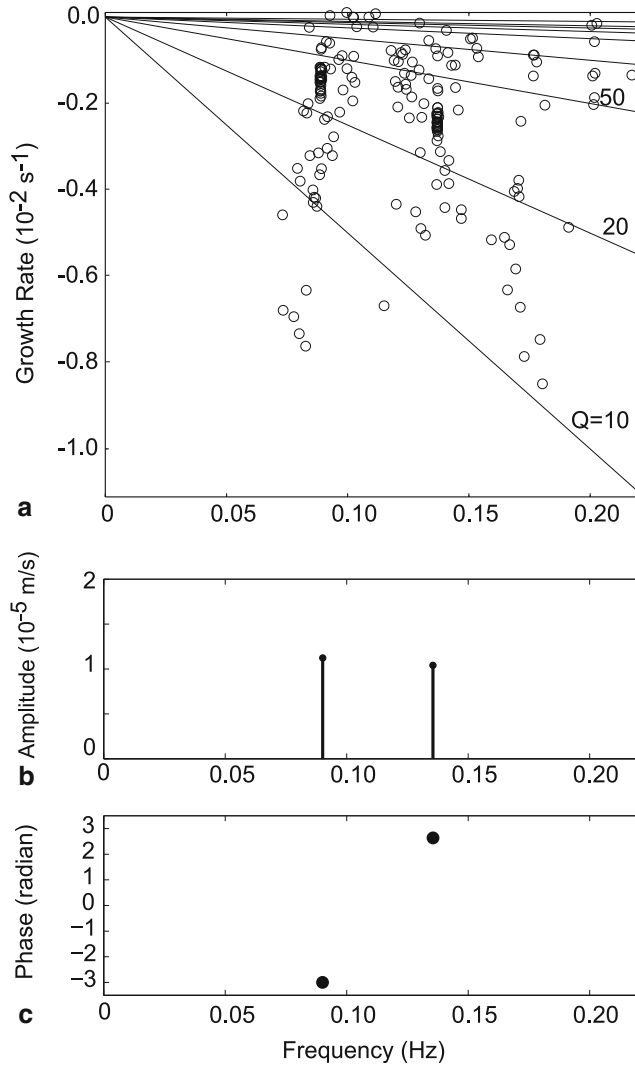
Some remarks on the theory of the Sompi method may be required, although Hori et al. (1989) and Kumazawa et al. (1990) stated that Sompi is a new spectral method. This may be incorrect, as suggested by Ulrych and Sacchi (1995). Let us consider again the AR Eq. (115), which is equivalently rewritten as

$$x(k) = \sum_{j=1}^{2m} a'(j)x(k-j), \quad (141)$$

where $a'(j) = -a(j)/a(0)$. From Eq. (114), we have

$$\begin{aligned} u(k) &= x(k) + e(k) \\ &= \sum_{j=1}^{2m} a'(j)x(k-j) + e(k). \end{aligned} \quad (142)$$

Substituting $u(k-j) = x(k-j) + e(k-j)$ into Eq. (142), we obtain



Source Quantification of Volcanic-Seismic Signals, Fig. 14 An example of the Sompi spectral analysis of the VLP waveform of Fig. 13a, in which Sompi was applied to the waveform after the time of the maximum amplitude indicated by the dashed line in Fig. 13a. (a) A frequency-growth rate diagram for estimated complex frequencies using trial AR orders between 4 and 60. Solid

lines represent lines along which the Q factor is constant. We can identify two densely populated regions, which represent the signal, while other scattered points represent incoherent noise. (b) Amplitudes and (c) phases at the time of the maximum amplitude using the two complex frequencies determined at an AR order of 40

$$u(k) = \sum_{j=1}^{2m} a'(j)u(k-j) + e(k) - \sum_{j=1}^{2m} a'(j)e(k-j). \quad (143)$$

This is the special autoregressive-moving average (ARMA) model discussed by Ulrych and Clayton (1976), who showed that the AR

coefficients in Eq. (143) can be determined by an eigenvalue problem that is essentially the same as that given in Eq. (120). In this sense, Sompi is not a new method. However, the Sompi theory, which is based not only on mathematics or information theory but also on physical concepts, yielded a systematic approach to estimate the complex frequencies and complex amplitudes, which can be directly related to the characteristic

properties of a linear dynamic system and the system excitation. This approach has been favorably used in analysis of LP and VLP events to estimate the complex frequencies and their temporal variations (e.g., De Angelis and McNutt 2005; Kumagai 2006; Kumagai and Chouet 1999; Kumagai et al. 2002a, 2003; Lin et al. 2005; Molina et al. 2004; Nakano and Kumagai 2005a; Nakano et al. 1998).

Fluid-Solid Interactions

Understanding the physics of the fluid-solid interactions is crucial to interpreting source mechanisms and spectral characters of volcano-seismic signals. The pioneering study on the fluid-solid interaction was by Biot (1952a), who studied seismic waves in a fluid-filled borehole and showed the existence of a slow wave propagating along the borehole boundary. This is called the tube wave. Aki et al. (1977) first studied a fluid-filled crack, although the motion of the fluid in the crack was not adequately treated in their study. Chouet (1986, 1988, 1992) fully studied the fluid-filled crack and demonstrated the existence of a slow wave in the fluid-filled crack, which was called the crack wave. We first consider a fluid layer sandwiched between two homogeneous elastic half-spaces following Ferrazzini and Aki (1987) to examine the physical properties of the crack wave. Then, we examine the resonance of a crack of finite

size containing a fluid by using the approach proposed by Chouet (1986). For simplicity, we assume an inviscid fluid.

Crack Wave

Let us consider a compressible fluid layer of thickness d , which is sandwiched between two elastic half-spaces bounded at $x_3 = -d/2$ and $x_3 = d/2$ (Fig. 15). The equation of motion for the fluid is given by

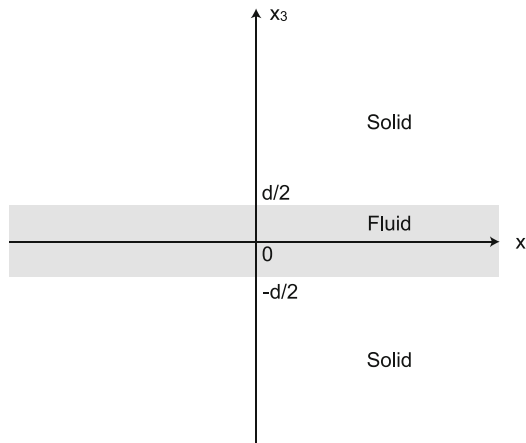
$$\rho_f \frac{\partial^2 u_1^f}{\partial t^2} = -\frac{\partial p}{\partial x_1}, \quad (144)$$

$$\rho_f \frac{\partial^2 u_3^f}{\partial t^2} = -\frac{\partial p}{\partial x_3}. \quad (145)$$

Here, ρ_f is the fluid density, u_1^f and u_3^f are the x_1 and x_3 components of the displacement in the fluid, respectively, and p is the fluid pressure satisfying the following continuity equation:

$$p = -b \left(\frac{\partial u_1^f}{\partial x_1} + \frac{\partial u_3^f}{\partial x_3} \right), \quad (146)$$

where b is the bulk modulus of the fluid. The equation of motion for the solid is



Source Quantification of Volcanic-Seismic Signals, Fig. 15 A fluid layer of thickness d sandwiched between two elastic half-spaces bounded at $x_3 = -d/2$ and $x_3 = d/2$

$$\rho_s \frac{\partial^2 u_1}{\partial t^2} = \frac{\partial \sigma_{11}}{\partial x_1} + \frac{\partial \sigma_{13}}{\partial x_3}, \quad (147)$$

$$\rho_s \frac{\partial^2 u_3}{\partial t^2} = \frac{\partial \sigma_{13}}{\partial x_1} + \frac{\partial \sigma_{33}}{\partial x_3}. \quad (148)$$

Here, ρ_s is the solid density, u_1 and u_3 are the x_1 and x_3 components of the displacement in the solid, respectively, and σ_{11} , σ_{33} , and σ_{13} are the components of stress:

$$\sigma_{11} = \lambda \left(\frac{\partial u_1}{\partial x_1} + \frac{\partial u_3}{\partial x_3} \right) + 2\mu \left(\frac{\partial u_1}{\partial x_1} \right), \quad (149)$$

$$\sigma_{33} = \lambda \left(\frac{\partial u_1}{\partial x_1} + \frac{\partial u_3}{\partial x_3} \right) + 2\mu \left(\frac{\partial u_3}{\partial x_3} \right), \quad (150)$$

$$\sigma_{13} = \mu \left(\frac{\partial u_1}{\partial x_3} + \frac{\partial u_3}{\partial x_1} \right). \quad (151)$$

The boundary conditions at the fluid-solid interfaces ($x_3 = \pm d/2$) are

$$u_3 = u_3^f, \quad (152)$$

$$\sigma_{33} = -p, \quad (153)$$

$$\sigma_{13} = 0. \quad (154)$$

The first and second conditions mean that the normal displacement and stress across the interface are both continuous. The third condition states that the shear stress vanishes at the interface.

For waves propagating in the x_1 direction, the displacements in the fluid may be written in the forms

$$u_1^f = r_1(x_3; k, \omega) e^{i(kx_1 - \omega t)}, \quad (155)$$

$$u_3^f = ir_2(x_3; k, \omega) e^{i(kx_1 - \omega t)}, \quad (156)$$

Substituting these equations into Eq. (146), we obtain

$$p = -ir_3 e^{i(kx_1 - \omega t)}, \quad (157)$$

where k and ω are the wavenumber and angular frequency, respectively

$$r_3 = b \left(kr_1 + \frac{dr_2}{dx_3} \right). \quad (158)$$

Substituting Eqs. (155) and (156) into Eqs. (144) and (145), respectively, and using Eq. (158), we have

$$\rho_f \omega^2 r_1 = kr_3, \quad (159)$$

$$-\rho_f \omega^2 r_2 = \frac{dr_3}{dx_3}. \quad (160)$$

Equations (158), (159), and (160) can be arranged in the following matrix form:

$$\begin{bmatrix} \frac{dr_2}{dx_3} \\ \frac{dr_3}{dx_3} \end{bmatrix} = \begin{bmatrix} 0 & \frac{1}{b} - \frac{k^2}{\rho_f \omega^2} \\ -\rho_f \omega^2 & 0 \end{bmatrix} \begin{bmatrix} r_2 \\ r_3 \end{bmatrix}. \quad (161)$$

For the solid, the displacement components of waves propagating in the x_1 direction may be written as

$$u_1 = y_1(x_3; k, \omega) e^{i(kx_1 - \omega t)}, \quad (162)$$

$$u_3 = iy_2(x_3; k, \omega) e^{i(kx_1 - \omega t)}. \quad (163)$$

Substituting these equations into Eqs. (147), (148), (149), (150), and (151), we obtain the relationship:

$$\begin{bmatrix} \frac{dy_1}{dx_3} \\ \frac{dy_2}{dx_3} \\ \frac{dy_3}{dx_3} \\ \frac{dy_4}{dx_3} \end{bmatrix} = \begin{bmatrix} 0 & k & \frac{1}{\mu} & 0 \\ \frac{-k\lambda}{\lambda+2\mu} & 0 & 0 & \frac{1}{\lambda+2\mu} \\ k^2\zeta - \rho_s \omega^2 & 0 & 0 & \frac{k\lambda}{\lambda+2\mu} \\ 0 & -\rho_s \omega^2 & -k & 0 \end{bmatrix} \begin{bmatrix} y_1 \\ y_2 \\ y_3 \\ y_4 \end{bmatrix}, \quad (164)$$

where $\zeta = \frac{4\mu(\lambda+\mu)}{\lambda+2\mu}$ and

$$y_3 = \mu \left(\frac{dy_1}{dx_3} - ky_2 \right), \quad (165)$$

$$y_4 = (\lambda + 2\mu) \frac{dy_2}{dx_3} + k\lambda y_1. \quad (166)$$

The boundary condition of Eqs. (152), (153), and (154) can then be written as

$$y_2 = r_2, \quad (167)$$

$$y_4 = r_3, \quad (168)$$

$$y_3 = 0, \quad (169)$$

respectively, at $x_3 = \pm d/2$. The systems of Eqs. (161) and (164) are to be solved under the boundary conditions of Eqs. (167), (168), and (169), in which we seek solutions satisfying the following two additional conditions

$$r_2(-x_3) = -r_2(x_3) \quad -d/2 \leq x_3 \leq d/2, \quad (170)$$

$$y_1, y_2 \rightarrow 0 \quad x_3 \rightarrow \pm\infty. \quad (171)$$

The first condition means that motion in the fluid layer is symmetric with respect to x_3 . There is the antisymmetric motion in the fluid layer with respect to x_3 . This motion is the same as that observed in a fluid layer overlying a solid half-space, such as ocean, and has been investigated by Biot (1952b) and Tolstoy (1954). This motion does not produce the slow wave, and I do not discuss it here.

The solution for the systems of Eqs. (161) and (164) under the conditions mentioned above can be determined by the following numerical procedure, which was proposed by Takeuchi and Saito (1972). Since we assume symmetric motion in the fluid layer, it is enough to consider the region $x_3 \geq 0$. We choose $x_3 = h$, where h is appropriately far from the fluid-solid interface to satisfy $y_1 = y_2 = 0$. At $x_3 = h$, we assume two independent initial values $y_1 = y_2 = y_4 = 0$, $y_3 = 1$ and

$y_1 = y_2 = y_3 = 0, y_4 = 1$ and integrate the equation for the solid (164) with values of k and ω by the Runge-Kutta method. Then, we have two independent solutions $y_i^A(x_3; k, \omega)$ and $y_i^B(x_3; k, \omega)$ ($i = 1, 2, 3, 4$). Any solution of Eq. (164) may be expressed as a linear combination of the two solutions:

$$y_i(x_3; k, \omega) = Ay_i^A(x_3; k, \omega) + By_i^B(x_3; k, \omega) \quad (i = 1, 2, 3, 4), \quad (172)$$

where A and B are arbitrary constants of integration for the two solutions. At $x_3 = 0$, we assume initial values $r_2 = 0$ and $r_3 = 1$ and integrate the equation for the fluid (161) with values of k and ω . Then, we have a solution $r_i^C(x_3; k, \omega)$ ($i = 2, 3$), and any solution of Eq. (161) may be given as

$$r_i(x_3; k, \omega) = Cr_i^C(x_3; k, \omega) \quad (i = 2, 3), \quad (173)$$

where C is an arbitrary constant of integration. From the boundary conditions of Eqs. (16), (168), and (169), we have the following relationships at $x_3 = d/2$:

$$\begin{aligned} Ay_2^A(d/2; k, \omega) + By_2^B(d/2; k, \omega) - Cr_2^C(d/2; k, \omega) &= 0, \\ Ay_4^A(d/2; k, \omega) + By_4^B(d/2; k, \omega) - Cr_3^C(d/2; k, \omega) &= 0, \\ Ay_3^A(d/2; k, \omega) + By_3^B(d/2; k, \omega) &= 0. \end{aligned} \quad (174)$$

These equations have a nontrivial solution if, and only if,

$$\det \begin{vmatrix} y_2^A(d/2; k, \omega) & y_2^B(d/2; k, \omega) & -r_2^C(d/2; k, \omega) \\ y_4^A(d/2; k, \omega) & y_4^B(d/2; k, \omega) & -r_3^C(d/2; k, \omega) \\ y_3^A(d/2; k, \omega) & y_3^B(d/2; k, \omega) & 0 \end{vmatrix} = 0. \quad (175)$$

The dispersion relation of the fluid-solid coupled waves can be determined by searching for k and ω that satisfy Eq. (175). From Eq. (174) we have

$$\frac{B}{A} = -\frac{y_3^A(d/2; k, \omega)}{y_3^B(d/2; k, \omega)}, \quad (176)$$

$$\frac{C}{A} = \frac{y_2^A(d/2; k, \omega)}{r_2^C(d/2; k, \omega)} - \frac{y_3^A(d/2; k, \omega)}{y_3^B(d/2; k, \omega)} \frac{y_2^B(d/2; k, \omega)}{r_2^C(d/2; k, \omega)}, \quad (177)$$

and the eigenfunctions are given as

$$y_i(x_3; k, \omega)/A = y_i^A(x_3; k, \omega) + (B/A)y_i^B(x_3; k, \omega) \quad (i = 1, 2, 3, 4), \quad (178)$$

$$r_i(x_3; k, \omega)/A = (C/A)r_i^C(x_3; k, \omega) \quad (i = 2, 3). \quad (179)$$

The dispersion relationship and eigenfunctions of the crack wave calculated by the above procedure are shown in Figs. 16 and 17, respectively. In this example, I used the following parameter values: $\rho_s = 1.10\rho_f$, $b = 0.06\mu$, and $\lambda = 2\mu$, which mimic a dike containing basaltic magma with a gas-volume fraction of 15%. The dispersion relationship in Fig. 16 clearly indicates that the phase velocity of the fluid-solid coupled wave depends on the wavelength, in which the velocity decreases as the wavelength increases. This wave is called the crack wave. Figure 17 displays dimensionless normal displacements (y_2 and r_2) and normal stress and pressure (y_4 and r_3) as a

function of dimensionless x_3 . Figure 17 indicates that the normal displacement is symmetric with respect to x_3 . This motion causes the deformation of the fluid-solid boundary. The deformation has the effect of lowering the phase velocity, which is examined below.

Fluid-Filled Crack Model

Chouet (1986, 1988, 1992) studied the motion of a finite crack containing a fluid in an infinite elastic medium. Following Chouet (1986), we consider a crack set in the plane $x_3 = 0$, which extends from $-W/2$ to $W/2$ along the x_1 axis and from 0 to L in the x_2 direction (Fig. 18). The thickness of the crack is d , which is much smaller than L . Chouet (1986) used the equations of motion for the solid given as

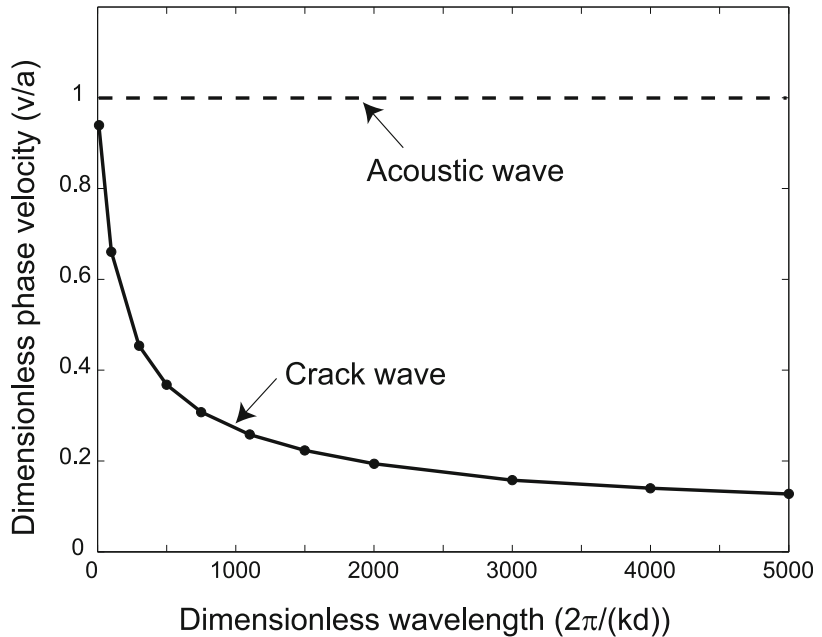
$$\rho_s \frac{\partial v_i}{\partial t} = \frac{\partial \sigma_{ij}}{\partial x_j}, \quad (180)$$

with the stress-strain relationship

$$\frac{\partial \sigma_{ij}}{\partial t} = \lambda \frac{\partial v_k}{\partial x_k} \delta_{ij} + \mu \left(\frac{\partial v_i}{\partial x_j} + \frac{\partial v_j}{\partial x_i} \right), \quad (181)$$

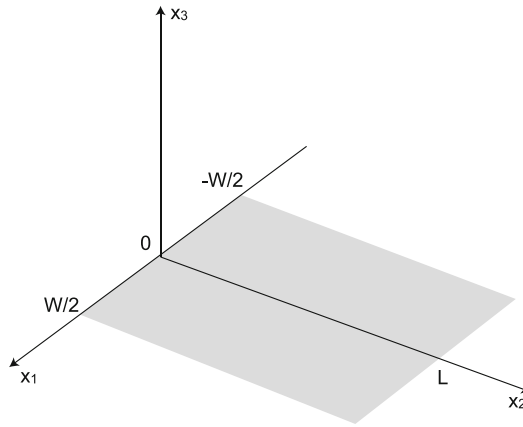
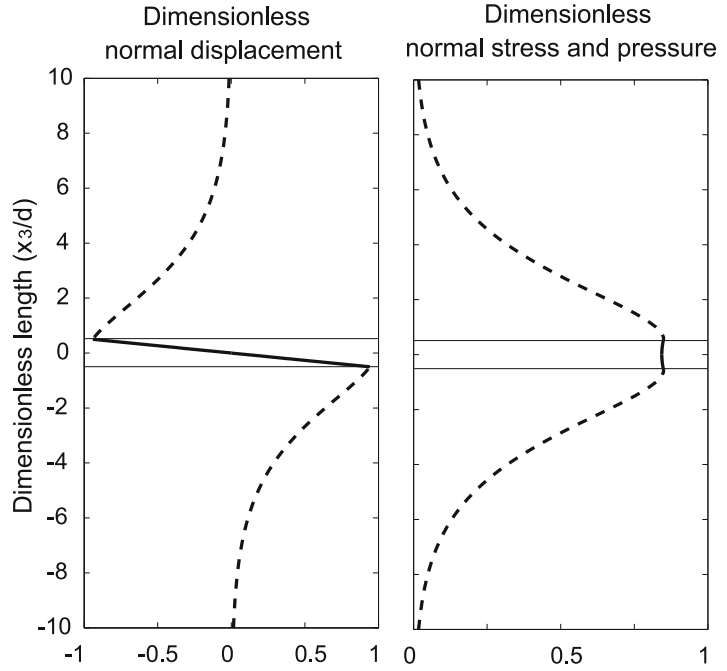
Source Quantification of Volcanic-Seismic Signals,

Fig. 16 The phase velocities of the acoustic wave (dashed line) and crack wave (solid line). These are normalized by the fluid acoustic velocity and plotted as a function of the wavelength normalized by the thickness of the fluid layer



Source Quantification of Volcanic-Seismic Signals,

Fig. 17 Dimensionless normal displacements, normal stress, and pressure as a function of x_3 normalized by the thickness of the fluid layer. *Thin solid lines* represent the boundaries between the fluid and solid, and *thick dashed and solid lines* represent the functions in the fluid and solid regions, respectively



Source Quantification of Volcanic-Seismic Signals, Fig. 18 Coordinates and geometry of a crack set in the plane $x_3 = 0$, which extends from $-W/2$ to $+W/2$ on the x_1 axis and from 0 to L on the x_2 axis

where v_j is particle velocity of the solid and $i = 1, 2, 3$. The equations of motion for the fluid are given as

$$\rho_f \frac{\partial^2 u_i^f}{\partial t^2} = -\frac{\partial p}{\partial x_i}, \quad (182)$$

with the equation of continuity

$$p = -b \frac{\partial u_k^f}{\partial x_k}. \quad (183)$$

The boundary conditions on the crack surface S_1 are

$$u_3^f = u_3, \quad \sigma_{33} = -p, \quad \sigma_{13} = \sigma_{23} = 0, \quad (184)$$

which are the continuity of normal displacement and stress and the vanishment of shear stress on S_1 . Let us introduce u_d , which represents the normal displacement of the crack wall and is identical to $u_3^f = u_3$ on S_1 .

We integrate Eq. (183) with respect to x_3 within the crack and divide it by d :

$$\frac{1}{d} \int_{-d/2}^{d/2} p dx_3 = -b \frac{1}{d} \int_{-d/2}^{d/2} \left[\frac{\partial u_1^f}{\partial x_1} + \frac{\partial u_2^f}{\partial x_2} + \frac{\partial u_3^f}{\partial x_3} \right] dx_3. \quad (185)$$

If we define

$$P = \frac{1}{d} \int_{-d/2}^{d/2} p dx_3 \quad (186)$$

and

$$U_i^f = \frac{1}{d} \int_{-d/2}^{d/2} u_i^f dx_3 \quad (i = 1, 2), \quad (187)$$

Eq. (185) is written as

$$P = -b \left(\frac{\partial U_1^f}{\partial x_1} + \frac{\partial U_2^f}{\partial x_2} + \frac{1}{d} \int_{-d/2}^{d/2} \frac{\partial u_3^f}{\partial x_3} dx_3 \right). \quad (188)$$

Since u_3^f is symmetric with respect to x_3 ($u_3^f(-x_3) = -u_3^f(x_3)$), we find that

$$\begin{aligned} \int_{-d/2}^{d/2} \frac{\partial u_3^f}{\partial x_3} dx_3 &= \int_{-d/2}^{d/2} du_3^f = [u_3^f]_{-d/2}^{d/2} \\ &= 2u_3^f(d/2). \end{aligned} \quad (189)$$

Note that $u_3^f(d/2) = u_3(d/2) = u_d$ from the boundary condition of Eq. (184) and, therefore, we have

$$P = -b \left(\frac{\partial U_1^f}{\partial x_1} + \frac{\partial U_2^f}{\partial x_2} + \frac{2}{d} u_d \right). \quad (190)$$

From Eq. (182) we obtain

$$\rho_f \frac{\partial^2 U_1^f}{\partial t^2} = -\frac{\partial P}{\partial x_1}, \quad (191)$$

$$\rho_f \frac{\partial^2 U_2^f}{\partial t^2} = -\frac{\partial P}{\partial x_2}. \quad (192)$$

Using the fluid velocity, Eqs. (191), (192), and (190) can be equivalently written as

$$\rho_f \frac{\partial V_1^f}{\partial t} = -\frac{\partial P}{\partial x_1}, \quad (193)$$

$$\rho_f \frac{\partial V_2^f}{\partial t} = -\frac{\partial P}{\partial x_2}, \quad (194)$$

$$\frac{\partial P}{\partial t} = -b \left(\frac{\partial V_1^f}{\partial x_1} + \frac{\partial V_2^f}{\partial x_2} + \frac{2}{d} v_d \right), \quad (195)$$

where $V_i^f = \partial U_i^f / \partial t$ ($i = 1, 2$) and v_d is the normal velocity of the crack wall.

Thus, the three-dimensional equations of motion and continuity are reduced to two-dimensional equations with respect to V_1^f, V_2^f, P , and v_d . In this problem, the boundary conditions are

$$\sigma_{33} = -p, \quad \sigma_{13} = \sigma_{23} = 0, \quad \text{on } S_1, \quad (196)$$

$$u_3 = 0, \quad \sigma_{13} = \sigma_{23} = 0, \quad \text{on } S_2, \quad (197)$$

where S_1 denotes the crack surface and S_2 denotes the crack plane outside the crack. The latter condition expresses that the displacement in the solid is symmetric with respect to x_3 . In addition to these boundary conditions, Chouet (1986) assumed zero mass transfer in and out of the crack, which are as follows:

$$V_1^f = v_1 \quad \text{along } |x_1| = W/2, \quad 0 \leq x_2 \leq L, \quad (198)$$

$$V_2^f = v_2 \quad \text{along } -W/2 \leq x_1 \leq W/2, \quad x_2 = 0, x_2 = L. \quad \text{or} \quad (199)$$

The equations of motion and stress-strain relationship for the solid in the form of Eqs. (180) and (181) and the equations of motion and continuity for the fluid in the forms of Eqs. (193), (194), and (195) are solved under the boundary conditions of Eqs. (196), (197), (198), and (199). Chouet (1986) solved this problem by using a finite-difference method, in which the following scaling was used:

$$\bar{x}_i = x_i/L, \quad (200)$$

$$\bar{t} = \alpha t/L, \quad (201)$$

$$\bar{\sigma}_{ij} = \sigma_{ij}/\sigma_0, \quad (202)$$

$$\bar{u}_i = u_i\mu/(L\sigma_0), \quad (203)$$

$$\bar{v}_i = v_i\mu/(\alpha\sigma_0), \quad (204)$$

where nondimensional variables are indicated by a bar, σ_0 is an effective stress, and α is the compressional wave velocity $\alpha = \sqrt{(\lambda + 2\mu)/\rho_s}$. Let us define the first-order difference operators in time and space as

$$\Delta_t f(l; i, j, k) = f(l+1; i, j, k) - f(l; i, j, k), \quad (205)$$

$$\Delta_{x1} f(l; i, j, k) = f(l; i+1, j, k) - f(l; i, j, k), \quad (206)$$

$$\Delta_{x2} f(l; i, j, k) = f(l; i, j+1, k) - f(l; i, j, k), \quad (207)$$

$$\Delta_{x3} f(l; i, j, k) = f(l; i, j, k+1) - f(l; i, j, k), \quad (208)$$

where $f(l; i, j, k)$ represents any function at time $l\Delta t$ and position $[i\Delta, j\Delta, k\Delta]$ or $[i\Delta_a, j\Delta_a, k\Delta_a]$. Here, Δt is the time increment and Δ and Δ_a are the grid spacings for the solid and fluid, respectively. A difference form of Eq. (195) is then obtained as

$$\frac{\Delta_t P}{\Delta t} = -b \left(\frac{\Delta_{x1} V_1^f}{\Delta_a} + \frac{\Delta_{x2} V_2^f}{\Delta_a} + \frac{2}{d} v_d \right) \quad (209)$$

$$\Delta_t P = -b H_a \left(\Delta_{x1} V_1^f + \Delta_{x2} V_2^f \right) - 2 \frac{b}{d} \Delta t v_d, \quad (210)$$

where $H_a = \Delta t/\Delta_a$. Using the scaling given by Eqs. (200) to (204), Eq. (210) can be further modified to a nondimensional form:

$$\Delta_t \bar{P} = - \left(\frac{b}{\mu} \right) \alpha H_a \left(\Delta_{x1} \bar{V}_1^f + \Delta_{x2} \bar{V}_2^f \right) - 2 C \Delta \bar{t} \bar{v}_d, \quad (211)$$

where $C = (b/\mu)(L/d)$ is called the crack stiffness (Aki et al. 1977). In a similar manner, we obtain nondimensional difference forms of equations of motion for the fluid (193) and (194) as

$$\Delta_t \bar{V}_1^f = - \left(\frac{a}{\alpha} \right)^2 \left(\frac{\mu}{b} \right) \alpha H_a \Delta_{x1} \bar{P}, \quad (212)$$

$$\Delta_t \bar{V}_2^f = - \left(\frac{a}{\alpha} \right)^2 \left(\frac{\mu}{b} \right) \alpha H_a \Delta_{x2} \bar{P}, \quad (213)$$

where $a = \sqrt{b/\rho_f}$ is the acoustic velocity of the fluid and the ratios α/a and b/μ are related to ρ_f and ρ_s through the relationship

$$\frac{\rho_f}{\rho_s} = \left(\frac{\mu}{\lambda + 2\mu} \right) \left(\frac{\alpha}{a} \right)^2 \left(\frac{b}{\mu} \right). \quad (214)$$

Nondimensional difference forms of the equations of motion and the stress-strain relationship for the solid (Eqs. (180) and (181)) are

$$\Delta_t \bar{v}_i = \frac{1}{2 + (\lambda/\mu)} \alpha H \Delta_{xi} \bar{\sigma}_{ij}, \quad (215)$$

$$\Delta_t \bar{\sigma}_{ij} = \alpha H \left[\frac{\lambda}{\mu} \Delta_{xk} \bar{v}_k \delta_{ij} + (\Delta_{xi} \bar{v}_j + \Delta_{xj} \bar{v}_i) \right], \quad (216)$$

where $H = \Delta t / \Delta$. To satisfy the conditions for numerical stability in the solid and fluid, Chouet (1986) used the following relationships:

$$H = \frac{\Delta t}{\Delta} = \frac{1}{4\alpha}, \quad (217)$$

$$H_a = \frac{\Delta t}{\Delta_a} = \frac{1}{2a}. \quad (218)$$

Applying a step increase in the fluid pressure over an area of the crack, the nondimensional difference equations for the solid and fluid are solved by a finite-difference method with a staggered grid scheme (Virieux 1986) to determine the normal velocity of the crack wall v_d . This source space-time function is used to synthesize the ground response to the fluid-filled crack. Note that Chouet (1986) added an artificial dissipation term to the equations for particle velocities in the solid (Eq. (215)) and for fluid velocities (Eqs. (212) and (213)) to suppress spurious numerical oscillations associated with discretization in time and space in the finite difference calculations.

Snapshots of the normal velocity of the crack wall v_d for a fluid-filled crack excited by a step increase in the pressure applied over a strip extending over the entire width of the crack are shown in Fig. 19a. The results show that the crack oscillation is dominated by the longitudinal modes with wavelengths L and $2L/5$. Figure 19b shows snapshots of v_d for the crack with the excitation applied over a small area at the center of the crack. The crack oscillation shown in Fig. 19b displays more complex spatial patterns than those of Fig. 19a, in which both the longitudinal and transverse modes with wavelengths $2L/3$ and $2W/3$, respectively, are dominantly excited. Chouet (1986) showed that the dimensionless phase velocity v/a of the crack wave sustaining the crack resonance depends on the crack stiffness C . The Q factor of the crack resonance almost monotonically increases with the increasing α/a (Kumagai and Chouet 2000, 2001).

The slow wave and its dependence on the crack stiffness may be understood by the following simple

theoretical consideration. Using Eqs. (190), (191), and (192), we obtain the following relationship:

$$\frac{\partial^2}{\partial t^2} \left(P + \frac{2b}{d} u_d \right) = \left(\frac{b}{\rho_f} \right) \left(\frac{\partial^2 P}{\partial x_1^2} + \frac{\partial^2 P}{\partial x_2^2} \right). \quad (219)$$

Note that $a = \sqrt{b/\rho_f}$ is the acoustic velocity of the fluid. Let us consider a simple case for which u_d is proportional to P :

$$u_d = \epsilon P, \quad (220)$$

where ϵ is a proportionality constant. Using the scaling given in Eqs. (202) and (203), the constant ϵ may be written in a nondimensional form as

$$\epsilon = \bar{\epsilon} L / \mu, \quad (221)$$

where $\bar{\epsilon}$ is a nondimensional proportionality constant. Using Eqs. (220) and (221), Eq. (219) can be written as

$$\left(1 + 2\bar{\epsilon} \frac{b}{\mu} \frac{L}{d} \right) \frac{\partial^2 P}{\partial t^2} = \left(\frac{b}{\rho_f} \right) \left(\frac{\partial^2 P}{\partial x_1^2} + \frac{\partial^2 P}{\partial x_2^2} \right) \quad (222)$$

or, equivalently,

$$\frac{\partial^2 P}{\partial t^2} = \left(\frac{b_e}{\rho_f} \right) \left(\frac{\partial^2 P}{\partial x_1^2} + \frac{\partial^2 P}{\partial x_2^2} \right), \quad (223)$$

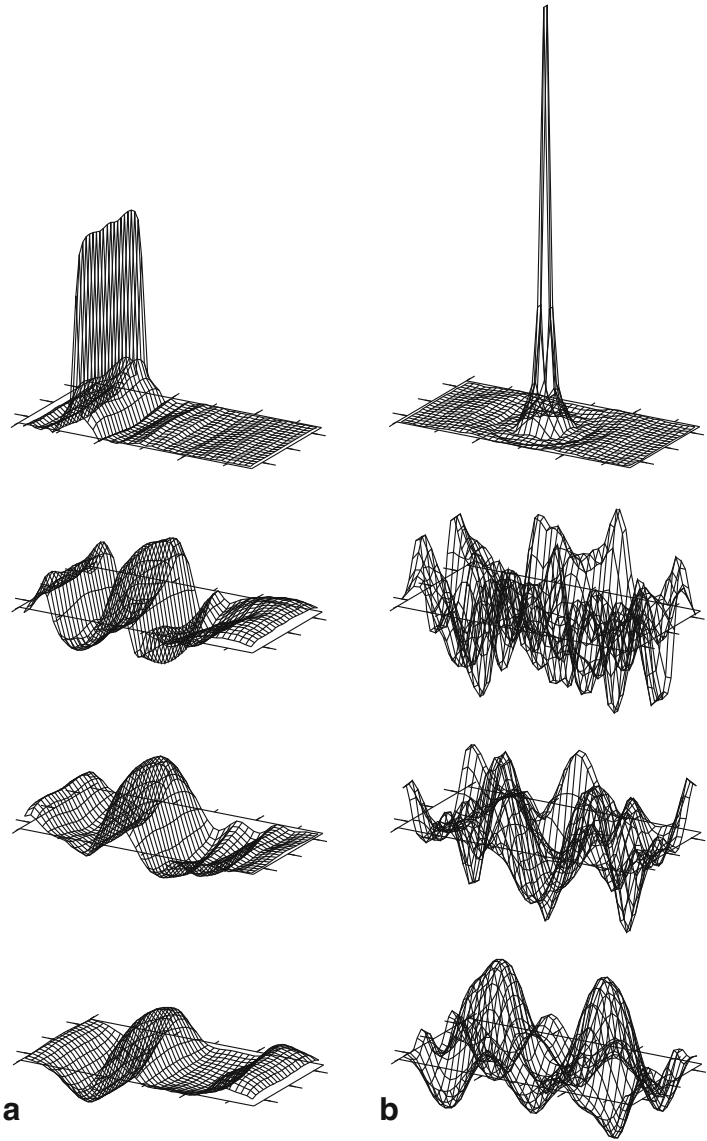
where b_e is an effective bulk modulus defined as

$$b_e = b / (1 + 2\bar{\epsilon} C) \quad (224)$$

Here, $C = (b/\mu)(L/d)$ is the crack stiffness. Equation (223) is the two-dimensional wave equation with phase velocity $a_e = \sqrt{b_e/\rho_f}$. If no deformation of the crack wall occurs ($\bar{\epsilon} = 0$), the phase velocity a_e is equal to the fluid acoustic velocity a . If the deformation of the crack wall occurs such that the wall moves inward with a decrease in fluid pressure ($\bar{\epsilon} > 0$), a_e becomes smaller than a through a reduction of the effective bulk modulus, which depends on the crack stiffness as given in Eq. (224).

Source Quantification of Volcanic-Seismic Signals,

Fig. 19 (a) Snapshots of the normal velocity of the crack wall v_d for a fluid-filled crack excited by a step increase in the pressure applied over a strip extending over the entire width of the crack. (b) Snapshots of v_d for the crack with the excitation applied over a small area at the center of the crack (Modified from Nakano et al. 2007)



The slow crack wave leads to more realistic estimates of the size and volume of a fluid-filled resonator as compared to a resonator with spherical geometry (e.g., Crosson and Bame 1985; Fujita and Ida 2003; Fujita et al. 1995; Sakuraba et al. 2002). The slow wave has been used to interpret spectral characteristics of LP and VLP events observed at various active volcanoes (Chouet et al. 1994; Gil Cruz and Chouet 1997; Jousset et al. 2003; Kumagai 2006; Kumagai and Chouet 1999, 2000; Kumagai et al. 2002a, 2003, 2005; Molina et al. 2004; Morrissey and Chouet 2001; Nakano and Kumagai 2005a;

Neuberg et al. 2000; Sturton and Neuberg 2006; Yamamoto 2005).

Future Directions

Recent volcano seismological studies have contributed greatly to achieving a better understanding of magmatic and hydrothermal systems. It is clear that the theoretical development of source models and analysis techniques has played an essential role in recent advances. The current studies in volcano seismology may

be summarized as follows: (a) physical understanding and modeling of source dynamics of LP and VLP events (Chouet et al. 2006; Neuberg et al. 2006; Nishimura 2004; Tuffen and Dingwell 2005), (b) development of a waveform inversion method of volcano-seismic signals for an extended source to achieve a better understanding of the resonance characteristics and triggering mechanisms (Nakano et al. 2007), (c) extension of the crack model to incorporate more realistic fluid acoustic properties (Kumagai 2006; Yamamoto 2005), and (d) dense broadband seismic observation and the use of waveform inversion techniques to monitor active volcanoes (e.g., Auger et al. 2006; Dawson et al. 2004; Kumagai et al. 2007). Future studies building on these will contribute further to our ability to quantify the sources of volcano-seismic signals, which will help us in our efforts to predict eruptive behavior and thus mitigate volcanic hazards through improved seismic monitoring of magmatic and hydrothermal activity.

Acknowledgments I am deeply grateful to Masaru Nakano for numerous discussions on all the subjects presented in this manuscript. I thank Yasuko Takei for constructive comments on the phenomenological source representation. Comments from Pablo Palacios, Luca D'Auria, Takeshi Nishimura, Yuta Maeda, and an anonymous reviewer helped improve the manuscript. I used the Generic Mapping Tools (GMT) (Wessel and Smith 1998) in the preparation of figures.

Green's Functions

Here, I briefly explain the relationship between the displacement and Green's functions. To simplify the explanation, I use two-dimensional equations in an infinite medium. The extension of the equations into three-dimension is straightforward. Green's functions defined by Eq. (5) are explicitly written as

$$\begin{aligned} \rho \frac{\partial^2 G_{11}}{\partial t^2} &= (\lambda + 2\mu) \frac{\partial^2 G_{11}}{\partial x_1^2} + (\lambda + \mu) \frac{\partial^2 G_{21}}{\partial x_2 \partial x_1} \\ &+ \mu \frac{\partial^2 G_{11}}{\partial x_2^2} + \delta(\mathbf{x} - \boldsymbol{\eta}) \delta(t - \tau), \end{aligned} \quad (225)$$

$$\begin{aligned} \rho \frac{\partial^2 G_{21}}{\partial t^2} &= (\lambda + 2\mu) \frac{\partial^2 G_{21}}{\partial x_2^2} \\ &+ (\lambda + \mu) \frac{\partial^2 G_{11}}{\partial x_2 \partial x_1} + \mu \frac{\partial^2 G_{21}}{\partial x_1^2}, \end{aligned} \quad (226)$$

$$\begin{aligned} \rho \frac{\partial^2 G_{12}}{\partial t^2} &= (\lambda + 2\mu) \frac{\partial^2 G_{12}}{\partial x_1^2} \\ &+ (\lambda + \mu) \frac{\partial^2 G_{22}}{\partial x_2 \partial x_1} + \mu \frac{\partial^2 G_{12}}{\partial x_2^2}, \end{aligned} \quad (227)$$

$$\begin{aligned} \rho \frac{\partial^2 G_{22}}{\partial t^2} &= (\lambda + 2\mu) \frac{\partial^2 G_{22}}{\partial x_2^2} + (\lambda + \mu) \frac{\partial^2 G_{12}}{\partial x_2 \partial x_1} \\ &+ \mu \frac{\partial^2 G_{22}}{\partial x_1^2} + \delta(\mathbf{x} - \boldsymbol{\eta}) \delta(t - \tau) \end{aligned} \quad (228)$$

Note that (G_{11}, G_{21}) represents the x_1 and x_2 components of the wavefield at $(\mathbf{x}, t)$, which is excited by the impulse in the x_1 direction applied at $\mathbf{x} = \boldsymbol{\eta}$ and $t = \tau$. Similarly, (G_{12}, G_{22}) represents the x_1 and x_2 components of the wavefield excited by the impulse in the x_2 direction at $\mathbf{x} = \boldsymbol{\eta}$ and at $t = \tau$. Therefore, i and j in G_{ij} represent the component and direction of the impulse, respectively. To specify the relationship between the receiver and source, we use the notation $G_{ij}(\mathbf{x}, t; \boldsymbol{\eta}, \tau)$. We obtain $G_{ij}(\mathbf{x}, t; \boldsymbol{\eta}, \tau) = G_{ij}(\mathbf{x}, t - \tau; \boldsymbol{\eta}, 0)$, since Green's functions are independent of the time of origin.

The displacement $u_i(\mathbf{x}, t)$ satisfies Eq. (3), which is explicitly written as

$$\begin{aligned} \rho \frac{\partial^2 u_1}{\partial t^2} &= (\lambda + 2\mu) \frac{\partial^2 u_1}{\partial x_1^2} + (\lambda + \mu) \frac{\partial^2 u_2}{\partial x_1 \partial x_2} \\ &+ \mu \frac{\partial^2 u_1}{\partial x_2^2} + f_1^S(\mathbf{x}, t), \end{aligned} \quad (229)$$

$$\begin{aligned} \rho \frac{\partial^2 u_2}{\partial t^2} &= (\lambda + 2\mu) \frac{\partial^2 u_2}{\partial x_2^2} + (\lambda + \mu) \frac{\partial^2 u_1}{\partial x_1 \partial x_2} \\ &+ \mu \frac{\partial^2 u_2}{\partial x_1^2} + f_2^S(\mathbf{x}, t) \end{aligned} \quad (230)$$

Equation (6) indicates that u_i is described by the following relationships:

$$u_1(\mathbf{x}, t) = \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} [f_1^S(\boldsymbol{\eta}, \tau) G_{11}(\mathbf{x}, t - \tau; \boldsymbol{\eta}, 0) + f_2^S(\boldsymbol{\eta}, \tau) G_{12}(\mathbf{x}, t - \tau; \boldsymbol{\eta}, 0)] d\boldsymbol{\eta}_1 d\boldsymbol{\eta}_2 d\tau, \quad (231)$$

$$u_2(\mathbf{x}, t) = \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} [f_1^S(\boldsymbol{\eta}, \tau) G_{21}(\mathbf{x}, t - \tau; \boldsymbol{\eta}, 0) + f_2^S(\boldsymbol{\eta}, \tau) G_{22}(\mathbf{x}, t - \tau; \boldsymbol{\eta}, 0)] d\boldsymbol{\eta}_1 d\boldsymbol{\eta}_2 d\tau. \quad (232)$$

We can verify that the displacement given by

$$L_1 = \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} \left[\rho \left(f_1^S \frac{\partial^2 G_{11}}{\partial t^2} + f_2^S \frac{\partial^2 G_{12}}{\partial t^2} \right) - (\lambda + 2\mu) \left(f_1^S \frac{\partial^2 G_{11}}{\partial x_1^2} + f_2^S \frac{\partial^2 G_{12}}{\partial x_1^2} \right) - (\lambda + \mu) \left(f_1^S \frac{\partial^2 G_{21}}{\partial x_1 \partial x_2} + f_2^S \frac{\partial^2 G_{22}}{\partial x_1 \partial x_2} \right) - \mu \left(f_1^S \frac{\partial^2 G_{11}}{\partial x_2^2} + f_2^S \frac{\partial^2 G_{12}}{\partial x_2^2} \right) \right] d\boldsymbol{\eta}_1 d\boldsymbol{\eta}_2 d\tau. \quad (234)$$

This equation can be modified as follows:

$$L_1 = \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} \left[f_1^S \left\{ \rho \frac{\partial^2 G_{11}}{\partial t^2} - (\lambda + 2\mu) \frac{\partial^2 G_{11}}{\partial x_1^2} - (\lambda + \mu) \frac{\partial^2 G_{21}}{\partial x_1 \partial x_2} - \mu \frac{\partial^2 G_{11}}{\partial x_2^2} \right\} + f_2^S \left\{ \rho \frac{\partial^2 G_{12}}{\partial t^2} - (\lambda + 2\mu) \frac{\partial^2 G_{12}}{\partial x_1^2} - (\lambda + \mu) \frac{\partial^2 G_{22}}{\partial x_1 \partial x_2} - \mu \frac{\partial^2 G_{12}}{\partial x_2^2} \right\} \right] d\boldsymbol{\eta}_1 d\boldsymbol{\eta}_2 d\tau. \quad (235)$$

From Eq. (227), we find that the second integral of Eq. (235) is zero. Therefore, Eq. (235) from Eq. (225) becomes

$$L_1 = \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} f_1^S(\boldsymbol{\eta}, \tau) \delta(\mathbf{x} - \boldsymbol{\eta}) \delta(t - \tau) d\boldsymbol{\eta}_1 d\boldsymbol{\eta}_2 d\tau = f_1^S(\mathbf{x}, t). \quad (236)$$

Similarly, we can show that the right-hand side of Eq. (233) is equivalent to $f_2^S(\mathbf{x}, t)$, and thus the displacement in the forms of Eqs. (231) and (232) satisfies the equation of motion (229) and (230).

Eqs. (231) and (232) satisfies Eqs. (229) and (230) in the following way. We can rewrite Eqs. (229) and (230) as

$$f_1^S(\mathbf{x}, t) = \rho \frac{\partial^2 u_1}{\partial t^2} - (\lambda + 2\mu) \frac{\partial^2 u_1}{\partial x_1^2} - (\lambda + \mu) \frac{\partial^2 u_2}{\partial x_1 \partial x_2} - \mu \frac{\partial^2 u_1}{\partial x_2^2}, \quad (233)$$

We denote the right-hand side of Eq. (233) as L_1 , which is derived from Eqs. (231) and (232) as

Moment Tensor for a Spherical Source

Let us consider the moment density tensors for three tensile cracks in the planes $\xi_1 = 0$, $\xi_2 = 0$, and $\xi_3 = 0$. If we sum and average these three tensors and assume that $[u_1] = [u_2] = [u_3] = D_s$, we obtain the following moment density tensor:

$$\mathbf{m} = D_s \begin{pmatrix} \lambda + 2\mu/3 & 0 & 0 \\ 0 & \lambda + 2\mu/3 & 0 \\ 0 & 0 & \lambda + 2\mu/3 \end{pmatrix}. \quad (237)$$

This represents the moment density tensor for the isotropic expansion of a cubic element.

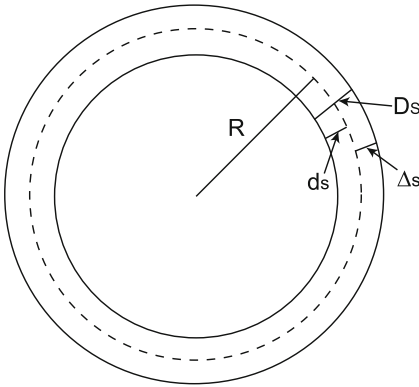
Following Müller (2001), we consider a spherical crack surface of radius R where a constant radial expansion $D_s = (d_s + \Delta_s)$ occurs (Fig. 20). Here, the inner wall of the crack moves inward by d_s and the outer wall moves outward by Δ_s . The moment tensor of the spherical expansion may be obtained by integration of the moment density tensor (237) over the surface R :

$$\mathbf{M} = \Delta V \begin{pmatrix} \lambda + 2\mu/3 & 0 & 0 \\ 0 & \lambda + 2\mu/3 & 0 \\ 0 & 0 & \lambda + 2\mu/3 \end{pmatrix}, \quad (238)$$

where ΔV is the volume given as

$$\Delta V = 4\pi R^2 D_s = 4\pi R^2 (d_s + \Delta_s) \quad (239)$$

The volume $4\pi R^2 d_s$ is caused by the inward motion, which compresses the sphere. The volume $\Delta V_s = 4\pi R^2 \Delta_s$, on the other hand, is caused by the outward motion, which excites seismic waves in the region outside the sphere. ΔV_s can be determined by solving an elastostatic boundary-value problem in the following way. We assume an isotropic medium and denote the



Source Quantification of Volcanic-Seismic Signals, Fig. 20 A spherical crack surface of radius R where a constant radial expansion $D_s = (d_s + \Delta_s)$ occurs. Here, the inner wall of the crack moves inward by d_s and the outer wall moves outward by Δ_s (Müller 2001)

regions inside and outside the sphere as regions 1 and 2, respectively. Since the motion is radial only, the equation of motion is given as (e.g., Takeuchi and Saito 1972)

$$\rho \frac{\partial^2 u}{\partial t^2} = \frac{1}{r^2} \frac{\partial}{\partial r} (r^2 \sigma_{rr}) - \frac{1}{r} (\sigma_{\theta\theta} + \sigma_{\phi\phi}) \quad (240)$$

and

$$\sigma_{rr} = (\lambda + 2\mu) \frac{\partial u}{\partial r} + \lambda \frac{2}{r} u, \quad (241)$$

$$\sigma_{\theta\theta} = \sigma_{\phi\phi} = \lambda \frac{\partial u}{\partial r} + (\lambda + \mu) \frac{2}{r} u, \quad (242)$$

where u is the radial displacement and σ_{rr} , $\sigma_{\theta\theta}$, and $\sigma_{\phi\phi}$ are the stress components in the spherical coordinate. Equations (240) and (241) apply to both regions 1 and 2. Substituting Eq. (241) into Eq. (240) and setting $\rho(\partial^2 u_r / \partial t^2) = 0$, we obtain the following static equilibrium equation:

$$\frac{\partial^2 u}{\partial r^2} + \frac{2}{r} \frac{\partial u}{\partial r} - \frac{2}{r^2} u = 0. \quad (243)$$

This equation has two solutions: $u = ar$ and $u = b/r^2$, where a and b are constants. The former is the interior solution for region 1 ($r \leq R$), and the latter is the exterior solution for region 2 ($r \geq R$). I denote the interior and exterior solutions as u_i and u_e , respectively. The constants a and b are determined by the boundary conditions for the radial displacement and the continuity of the radial stress at $r = R$:

$$u_e(R) - u_i(R) = \frac{b}{R^2} - aR = D_s, \quad (244)$$

$$\begin{aligned} \sigma_{rr}^e(R) - \sigma_{rr}^i(R) &= -\frac{4\mu}{R^3} b - (3\lambda + 2\mu)a \\ &= 0, \end{aligned} \quad (245)$$

where σ_{rr}^i and σ_{rr}^e are the radial stresses in regions 1 and 2, respectively. Accordingly, we obtain

$$u_i(r) = -\frac{4\mu D_s}{3(\lambda + 2\mu)} \frac{r}{R} \quad (r \leq R), \quad (246)$$

$$u_e(r) = \frac{(\lambda + 2\mu/3)D_s}{\lambda + 2\mu} \frac{R^2}{r^2} \quad (r \geq R) \quad (247)$$

Then, we obtain

$$d_s = -u_i(R) = \frac{4\mu D_s}{3(\lambda + 2\mu)}, \quad (248)$$

$$\Delta_s = u_e(R) = \frac{(\lambda + 2\mu/3)D_s}{\lambda + 2\mu}. \quad (249)$$

Equation (239) can be modified as

$$\Delta V = 4\pi R^2 \Delta_s (D_s / \Delta_s) = \frac{\lambda + 2\mu}{\lambda + 2\mu/3} \Delta V_s. \quad (250)$$

Finally, we obtain the moment tensor for the spherical expansion as

$$\mathbf{M} = (\lambda + 2\mu) \Delta V_s \begin{pmatrix} 1 & 0 & 0 \\ 0 & 1 & 0 \\ 0 & 0 & 1 \end{pmatrix}. \quad (251)$$

Moment Tensor for a Cylindrical Source

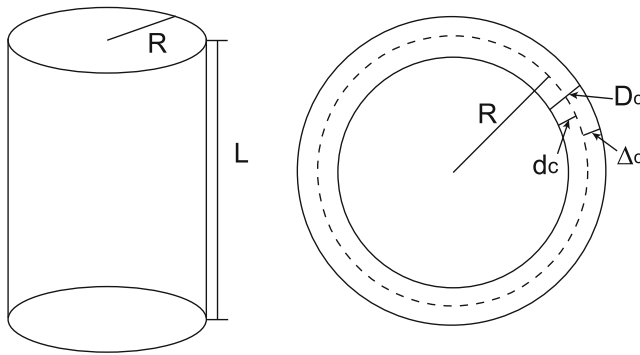
We consider the moment density tensors for two tensile cracks in the planes $\xi_1 = 0$ and $\xi_2 = 0$, where ξ_1 and ξ_2 are two horizontal axes. If we sum and average these two tensors and assume that $[u_1] = [u_2] = D_c$, we obtain

$$\mathbf{m} = D_c \begin{pmatrix} \lambda + \mu & 0 & 0 \\ 0 & \lambda + \mu & 0 \\ 0 & 0 & \lambda \end{pmatrix}. \quad (252)$$

This represents the moment density tensor for the expansion of a cubic element in the two horizontal directions. Let us consider a vertical cylinder of length L and radius R . The cylinder surface at radius R can be regarded as a cylindrical crack, where the radial expansion $D_c = (d_c + \Delta_c)$ occurs (Fig. 21). The moment tensor for the radial expansion of the cylinder may be obtained by integration of the moment density tensor (252) over the surface R :

$$\mathbf{M} = \Delta V \begin{pmatrix} \lambda + \mu & 0 & 0 \\ 0 & \lambda + \mu & 0 \\ 0 & 0 & \lambda \end{pmatrix}, \quad (253)$$

where ΔV is the volume given as



Source Quantification of Volcanic-Seismic Signals, Fig. 21 A vertical cylinder of length L and radius R . The cylinder surface can be regarded as a cylindrical crack,

where the radial expansion $D_c = (d_c + \Delta_c)$ occurs. Here, the inner wall of the crack moves inward by d_c and the outer wall moves outward by Δ_c (Müller 2001)

$$\Delta V = 2\pi R L D_c = 2\pi R L (d_c + \Delta_c) \quad (254)$$

We obtain the static equilibrium equation for the radial expansion of the cylinder u as

$$\frac{\partial^2 u}{\partial r^2} + \frac{1}{r} \frac{\partial u}{\partial r} - \frac{1}{r^2} u = 0. \quad (255)$$

This equation has internal and external solutions, which are given as $u = ar$ and $u = b/r$, respectively. The constants a and b are determined by the boundary conditions

$$u_e(R) - u_i(R) = \frac{b}{R} - aR = D_c, \quad (256)$$

$$\begin{aligned} \sigma_{rr}^e(R) - \sigma_{rr}^i(R) &= -\frac{2\mu}{R^2} b - 2(\lambda + \mu)a \\ &= 0, \end{aligned} \quad (257)$$

where superscripts i and e denote the interior and exterior solutions. We then obtain

$$d_c = -u_i(R) = \frac{\mu D_s}{(\lambda + 2\mu)}, \quad (258)$$

$$\Delta_c = u_e(R) = \frac{(\lambda + \mu) D_c}{\lambda + 2\mu}, \quad (259)$$

and

$$\Delta V = \frac{\lambda + 2\mu}{\lambda + \mu} \Delta V_c. \quad (260)$$

The moment tensor for the vertical cylinder is therefore given as

$$\mathbf{M} = \frac{\lambda + 2\mu}{\lambda + \mu} \Delta V_c \begin{pmatrix} \lambda + \mu & 0 & 0 \\ 0 & \lambda + \mu & 0 \\ 0 & 0 & \lambda \end{pmatrix}. \quad (261)$$

Let us rotate the vertical cylinder with axis orientation angles ϕ and θ (Fig. 8b). This can be done in the following steps: (1) fix the cylinder, (2) rotate the ξ_1 and ξ_2 axes around the ξ_3 axis through an angle $-\phi$, and (3) further rotate the ξ_1 and ξ_3 axes around the ξ_2 axis through an angle $-\theta$. The rotation matrix $\mathbf{R}$ is given as

$$\mathbf{R} = \begin{pmatrix} \cos \theta & 0 & -\sin \theta \\ 0 & 1 & 0 \\ \sin \theta & 0 & \cos \theta \end{pmatrix} \begin{pmatrix} \cos \phi & -\sin \phi & 0 \\ \sin \phi & \cos \phi & 0 \\ 0 & 0 & 1 \end{pmatrix} \quad (262)$$

$$\mathbf{R} = \begin{pmatrix} \cos \theta \cos \phi & -\cos \theta \sin \phi & -\sin \theta \\ \sin \phi & \cos \phi & 0 \\ \sin \theta & -\sin \theta \sin \phi & \cos \theta \end{pmatrix}. \quad (263)$$

Using the matrix $\mathbf{R}$, we obtain the moment tensor for a cylinder with the axis orientation angles (ϕ, θ) as

$$\mathbf{M}' = \mathbf{R}^T \mathbf{M} \mathbf{R}, \quad (264)$$

which leads to Eq. (34).

Bibliography

Primary Literature

- Aki K, Richards PG (2002) Quantitative seismology, 2nd edn. University Science Books, Sausalito
- Aki K, Fehler M, Das S (1977) Source mechanism of volcanic tremor: fluid-driven crack models and their application to the 1963 Kilauea eruption. *J Volcanol Geotherm Res* 2:259–287
- Almendros J, Chouet BA, Dawson PB (2001a) Spatial extent of a hydrothermal system at Kilauea volcano, Hawaii, determined from array analyses of shallow long-period seismicity 1. Method. *J Geophys Res* 106:13565–13580
- Almendros J, Chouet BA, Dawson PB (2001b) Spatial extent of a hydrothermal system at Kilauea volcano, Hawaii, determined from array analyses of shallow long-period seismicity 2. Results. *J Geophys Res* 106:13581–13597
- Almendros J, Chouet BA, Dawson PB, Huber C (2002a) Mapping the sources of the seismic wave field at Kilauea volcano, Hawaii, using data recorded on multiple seismic antennas. *Bull Seism Soc Am* 92:2333–2351
- Almendros J, Chouet BA, Dawson PB, Bond T (2002b) Identifying elements of the plumbing system beneath Kilauea volcano, Hawaii, from the source locations of very-long-period signals. *Geophys J Int* 148:303–312
- Aoyama H, Takeo M (2001) Wave properties and focal mechanisms of N-type earthquakes at Asama volcano. *J Volcanol Geotherm Res* 105:163–182

- Arciniega-Ceballos A, Chouet BA, Dawson PB (1999) Very long-period signals associated with vulcanian explosions at Popocatepetl volcano, Mexico. *Geophys Res Lett* 26:3013–3016
- Arciniega-Ceballos A, Chouet BA, Dawson PB (2003) Long-period events and tremor at Popocatepetl volcano (1994–2000) and their broadband characteristics. *Bull Volcanol* 65:124–135
- Waster R, Mah S, Kyle P, McIntosh W, Dunbar N, Johnson J, Ruiz M, McNamara S (2003) Very long period oscillations of mount Erebus Volcano. *J Geophys Res* 108:2522. <https://doi.org/10.1029/2002JB002101>
- Auger E, D'Auria L, Martini M, Chouet BA, Dawson PB (2006) Real-time monitoring and massive inversion of source parameters of very-long-period (VLP) seismic signals: an application to Stromboli Volcano, Italy. *Geophys Res Lett* 33:L04301. <https://doi.org/10.1029/2005GL024703>
- Backus G, Mulcahy M (1976) Moment tensors and other phenomenological descriptions of seismic sources-I. Continuous displacements. *Geophys J R Astr Soc* 46:341–361
- Battaglia J, Aki K (2003) Location of seismic events and eruptive fissures on the Piton de la Fournaise volcano using seismic amplitudes. *J Geophys Res* 108:2364. <https://doi.org/10.1029/2002JB002193>
- Biot MA (1952a) Propagation of elastic waves in a cylindrical bore containing a fluid. *J Appl Phys* 23:997–1005
- Biot MA (1952b) The interaction of Rayleigh and Stoneley waves in the ocean bottom. *Bull Seism Soc Am* 42:82–92
- Bouchon M (1979) Discrete wave number representation of elastic wave fields in three-space dimensions. *J Geophys Res* 84:3609–3614
- Bouchon M (1981) A simple method to calculate Green's functions for elastic layered media. *Bull Seism Soc Am* 71:959–971
- Bouchon M, Schultz CA, Toksöz MN (1996) Effect of 3D topography on seismic motion. *J Geophys Res* 101:5835–5846
- Brandstøttir B, Einarsson P (1979) Seismic activity associated with the September 1977 deflation of the Krafla central volcano in Northeastern Iceland. *J Volcanol Geotherm Res* 6:197–212
- Chouet BA (1982) Free surface displacements in the near field of a tensile crack expanding in three dimensions. *J Geophys Res* 87:3868–3872
- Chouet BA (1985) Excitation of a buried magmatic pipe: a seismic source model for volcanic tremor. *J Geophys Res* 90:1881–1893
- Chouet BA (1986) Dynamics of a fluid-driven crack in three dimensions by the finite difference method. *J Geophys Res* 91:13967–13992
- Chouet BA (1988) Resonance of a fluid-driven crack: radiation properties and implications for the source of long-period events and harmonic tremor. *J Geophys Res* 93:4375–4400
- Chouet BA (1992) A seismic model for the source of long-period events and harmonic tremor. In: Gasparini P, Scarpa R, Aki K (eds) *Volcanic seismology*. Springer, Berlin, pp 133–156
- Chouet BA (1996a) New methods and future trends in seismological volcano monitoring. In: Scarpa R, Tilling RI (eds) *Monitoring and mitigation of volcano hazards*. Springer, New York, pp 23–97
- Chouet BA (1996b) Long-period volcano seismicity: its source and use in eruption forecasting. *Nature* 380:309–316
- Chouet BA (2003) *Volcano Seismology*. Pure Appl Geophys 160:739–788
- Chouet BA, Julian B (1985) Dynamics of an expanding fluid-filled crack. *J Geophys Res* 90:11187–11198
- Chouet BA, Page RA, Stephens CD, Lahr JC, Power JA (1994) Precursory swarms of long-period events at redoubt volcano (1989–1990), Alaska: their origin and use as a forecasting tool. *J Volcanol Geotherm Res* 62:95–135
- Chouet BA, Saccorotti G, Martini M, Dawson PB, De Luca G, Milana G, Scarpa R (1997) Source and path effects in the wavefields of tremor and explosions at Stromboli volcano, Italy. *J Geophys Res* 102:15129–15150
- Chouet BA, Saccorotti G, Dawson PB, Martini M, Scarpa R, De Luca G, Milana G, Cattaneo M (1999) Broadband measurements of the sources of explosions at Stromboli Volcano, Italy. *Geophys Res Lett* 26:1937–1940
- Chouet BA, Dawson PB, Ohminato T, Martini M, Saccorotti G, Guidicepierto F, De Luca G, Milana G, Scarpa R (2003) Source mechanisms of explosions at Stromboli Volcano, Italy, determined from moment-tensor inversions of very-long-period data. *J Geophys Res* 108:2331. <https://doi.org/10.1029/2003JB002535>
- Chouet BA, Dawson PB, Arciniega-Ceballos A (2005) Source mechanism of Vulcanian degassing at Popocatepetl volcano, Mexico, determined from waveform inversions of very long period signals. *J Geophys Res* 110:B07301. <https://doi.org/10.1029/2004JB003524>
- Chouet BA, Dawson PB, Nakano M (2006) Dynamics of diffusive bubble growth and pressure recovery in a bubbly rhyolitic melt embedded in an elastic solid. *J Geophys Res* 111:B07310. <https://doi.org/10.1029/2005JB004174>
- Crosson RS, Bame DA (1985) A spherical source model for low frequency volcanic earthquakes. *J Geophys Res* 90:10237–10247
- Dawson PB, Dietel C, Chouet BA, Honma K, Ohminato T, Okubo P (1998) A digitally telemetered broadband seismic network at Kilauea volcano, Hawaii. *US Geol Surv Open File Rep* 98(108):1–121
- Dawson PB, Whilldin D, Chouet BA (2004) Application of near real-time semblance to locate the shallow magmatic conduit at Kilauea volcano, Hawaii. *Geophys Res Lett* 31:L21606. <https://doi.org/10.1029/2004GL021163>
- De Angelis S, McNutt SR (2005) Degassing and hydrothermal activity at Mt. Spurr, Alaska during the summer

- of 2004 inferred from the complex frequencies of long-period events. *Geophys Res Lett* 32:L12312. <https://doi.org/10.1029/2005GL022618>
- Ferrazzini V, Aki K (1987) Slow waves trapped in a fluid-filled infinite crack: implication for volcanic tremor. *J Geophys Res* 92:9215–9223
- Fujita E, Ida Y (2003) Geometrical effects and low-attenuation resonance of volcanic fluid inclusions for the source mechanism of long-period earthquakes. *J Geophys Res* 108:2118. <https://doi.org/10.1029/2002JB001806>
- Fujita E, Ida Y, Oikawa J (1995) Eigen oscillation of a fluid sphere and source mechanism of harmonic volcanic tremor. *J Volcanol Geotherm Res* 69:365–378
- Fujita E, Ukawa M, Yamamoto E (2004) Subsurface cyclic magma sill expansions in the 2000 Miyakejima volcano eruption: possibility of two-phase flow oscillation. *J Geophys Res* 109:B04205. <https://doi.org/10.1029/2003JB002556>
- Fukuyama E, Kubo A, Kawai H, Nonomura K (2001) Seismic remote monitoring of stress field. *Earth Planets Space* 53:1021–1026
- Gil Cruz F, Chouet BA (1997) Long-period events, the most characteristic seismicity accompanying the emplacement and extrusion of a lava dome in Galeras volcano, Colombia, in 1991. *J Volcanol Geotherm Res* 77:121–158
- Goldstein P, Chouet BA (1994) Array measurements and modeling of sources of shallow volcanic tremor at Kilauea volcano, Hawaii. *J Geophys Res* 99:2637–2652
- Hayashi Y, Morita Y (2003) An image of a magma intrusion process inferred from precise hypocenter migrations of the earthquake swarm east of the Izu Peninsula. *Geophys J Int* 153:159–174
- Hidayat D, Voight B, Langston C, Ratdomopurbo A, Ebeling C (2000) Broadband seismic experiment at Merapi volcano, Java, Indonesia: very-long-period pulses embedded in multiphase earthquakes. *J Volcanol Geotherm Res* 100:215–231
- Hidayat D, Chouet BA, Voght B, Dawson P, Ratdomopurbo A (2002) Source mechanism of very-long-period signals accompanying dome growth activity at Merapi volcano, Indonesia. *Geophys Res Lett* 23:2118. <https://doi.org/10.1029/2002GL015013>
- Hill DP, Dawson PB, Johnston MJS, Pitt AM (2002) Very-long-period volcanic earthquakes beneath Mammoth Mountain, California. *Geophys Res Lett* 29:1370. <https://doi.org/10.1029/2002GL014833>
- Hori S, Fukao Y, Kumazawa M, Furumoto M, Yamamoto A (1989) A new method of spectral analysis and its application to the earth's free oscillations: the “Sompi” method. *J Geophys Res* 94:7535–7553
- Iguchi M (1994) A vertical expansion source model for the mechanisms of earthquakes originated in the magma conduit of an andesitic volcano: Sakurajima, Japan. *Bull Volcanol Soc Jpn* 39:49–67
- Jousset P, Neuberg JW, Sturton S (2003) Modelling the time-dependent frequency content of low-frequency volcanic earthquakes. *J Volcanol Geotherm Res* 128:201–223
- Julian BR (1994) Volcanic tremor: nonlinear excitation by fluid flow. *J Geophys Res* 99:11859–11877
- Julian BR, Miller AD, Foulger GR (1998) Non-double-couple earthquakes 1. Theory. *Rev Geophys* 36:525–549
- Kanamori H, Given J (1982) Analysis of long-period seismic waves excited by the May 18, 1980, eruption of Mount St. Helens – a terrestrial monopole. *J Geophys Res* 87:5422–5432
- Kanamori H, Given J, Lay T (1984) Analysis of seismic body waves excited by the Mount St. Helens eruption of May 18, 1980. *J Geophys Res* 89:1856–1866
- Kaneshima S et al (1996) Mechanism of phreatic eruptions at Aso volcano inferred from near-field broadband observations. *Science* 273:642–645
- Kawakatsu H, Ohminato T, Ito H, Kuwahara Y, Kato T, Tsuruga K, Honda S, Yomogida K (1992) Broadband seismic observation at Sakurajima volcano, Japan. *Geophys Res Lett* 19:1959–1962
- Kawakatsu H, Ohminato T, Ito H (1994) 10s-period volcanic tremors observed over a wide area in southwestern Japan. *Geophys Res Lett* 21:1963–1966
- Kawakatsu H et al (2000) Aso-94: Aso seismic observation with broadband instruments. *J Volcanol Geotherm Res* 101:129–154
- Kawakatsu H, Yamamoto M (2007) Volcano Seismology. In: Kamanori H (ed) *Treatise on geophysics. Earthquake seismology*, vol 4. Elsevier, Amsterdam, pp 389–420
- Kawase H (1988) Time domain response of a semi-circular canyon for incident SV, P and Rayleigh waves calculated by the discrete wavenumber boundary element method. *Bull Seism Soc Am* 78:1415–1437
- Kennet LN, Kerry NJ (1979) Seismic waves in a stratified half-space. *Geophys J R Astr Soc* 57:557–583
- Kobayashi T, Ohminato T, Ida Y (2003) Earthquakes series preceding very long period seismic signals observed during the 2000 Miyakejima volcanic activity. *Geophys Res Lett* 30:1423. <https://doi.org/10.1029/2002GL016631>
- Kumagai H (2006) Temporal evolution of a magmatic dike system inferred from the complex frequencies of very long period seismic events. *J Geophys Res* 111: B06201. <https://doi.org/10.1029/2005JB003881>
- Kumagai H, Chouet BA (1999) The complex frequencies of long-period seismic events as probes of fluid composition beneath volcanoes. *Geophys J Int* 138:F7–F12
- Kumagai H, Chouet BA (2000) Acoustic properties of a crack containing magmatic or hydrothermal fluids. *J Geophys Res* 105:25493–25512
- Kumagai H, Chouet BA (2001) The dependence of acoustic properties of a crack on the mode and geometry. *Geophys Res Lett* 28:3325–3328
- Kumagai H, Ohminato T, Nakano M, Ooi M, Kubo A, Inoue H, Oikawa J (2001) Very-long-period seismic signals and caldera formation at Miyake Island. *Jpn Sci* 293:687–690

- Kumagai H, Chouet BA, Nakano M (2002a) Temporal evolution of a hydrothermal system in Kusatsu-Shirane volcano, Japan, inferred from the complex frequencies of long-period events. *J Geophys Res* 107:2236. <https://doi.org/10.1029/2001JB000653>
- Kumagai H, Chouet BA, Nakano M (2002b) Waveform inversion of oscillatory signatures in long-period events beneath volcanoes. *J Geophys Res* 107:2301. <https://doi.org/10.1029/2001JB001704>
- Kumagai H, Miyakawa K, Negishi H, Inoue H, Obara K, Suetsugu D (2003) Magmatic dike resonances inferred from very-long-period seismic signals. *Science* 299:2058–2061. <https://doi.org/10.1126/science.1081195>
- Kumagai H, Chouet BA, Dawson PB (2005) Source process of a long-period event at Kilauea volcano, Hawaii. *Geophys J Int* 161:243–254
- Kumagai H et al (2007) Enhancing volcano-monitoring capabilities in Ecuador. *Eos Trans AGU* 88:245–246
- Kumazawa M, Imanishi Y, Fukao Y, Furumoto M, Yamamoto A (1990) A theory of spectral analysis based on the characteristic property of a linear dynamic system. *Geophys J Int* 101:613–630
- Legrand D, Kaneshima S, Kawakatsu H (2000) Moment tensor analysis of near field broadband waveforms observed at Aso volcano, Japan. *J Volcanol Geotherm Res* 101:155–169
- Lesage P, Glangaud F, Mars J (2002) Applications of autoregressive models and time-frequency analysis to the study of volcanic tremor and long-period events. *J Volcanol Geotherm Res* 114:391–417
- Lin CH, Konstantinou KI, Liang WT, Pu HC, Lin YM, You SH, Huang YP (2005) Preliminary analysis of volcanoseismic signals recorded at the Tatun volcano group, Northern Taiwan. *Geophys Res Lett* 32:L10313. <https://doi.org/10.1029/2005GL022861>
- Matsuura T, Imanishi Y, Imanari M, Kumazawa M (1990) Application of a new method of high-resolution spectral analysis, “Sompri”, for free induction decay of nuclear magnetic resonance. *Appl Spectrosc* 44:618–626
- McNutt SR (1996) Seismic monitoring and eruption forecasting of volcanoes: a review of the state-of-the art and cased histories. In: Scarpa R, Tilling RI (eds) *Monitoring and mitigation of volcano hazards*. Springer, New York, pp 99–146
- McNutt SR (2005) Volcanic seismology. *Annu Rev Earth Planet Sci* 33:461–491
- Métaxian JP, Lesage P, Dorel J (1997) Permanent tremor of Masaya volcano, Nicaragua: wavefield analysis and source location. *J Geophys Res* 102:22529–22545
- Menke W (1989) *Geophysical data analysis: discrete inverse theory*, Revised edn. Academic, San Diego
- Minakami T (1974) Seismology of volcanoes in Japan. In: Civetta L, Gasparini P, Luongo G, Rapolla A (eds) *Physical volcanology*. Elsevier, Amsterdam, pp 1–27
- Molina I, Kumagai H, Yepes H (2004) Resonances of a volcanic conduit triggered by repetitive injections of an ash-laden gas. *Geophys Res Lett* 31:L03603. <https://doi.org/10.1029/2003GL018934>
- Morita Y, Nakao S, Hayashi Y (2006) A quantitative approach to the dike intrusion process inferred from a joint analysis of geodetic and seismological data for the 1998 earthquake swarm off the east coast of Izu Peninsula, Central Japan. *J Geophys Res* 111:B06208. <https://doi.org/10.1029/2005JB003860>
- Morrissey M, Chouet BA (1997) A numerical investigation of choked flow dynamics and its application to the triggering mechanism of long-period events at redoubt volcano, Alaska. *J Geophys Res* 102:7965–7983
- Morrissey M, Chouet BA (2001) Trends in long-period seismicity related to magmatic fluid compositions. *J Volcanol Geotherm Res* 108:265–281
- Müller G (2001) Volume change of seismic sources from moment tensors. *Bull Seism Soc Am* 91:880–884
- Nakamichi H, Hamaguchi H, Tanaka S, Ueki S, Nishimura T, Hasegawa A (2003) Source mechanisms of deep and intermediate-depth low-frequency earthquakes beneath Iwate volcano, Northeastern Japan. *Geophys J Int* 154:811–828
- Nakano M, Kumagai H (2005a) Response of a hydrothermal system to magmatic heat inferred from temporal variations in the complex frequencies of long-period events at Kusatsu-Shirane volcano, Japan. *J Volcanol Geotherm Res* 147:233–244
- Nakano M, Kumagai H (2005b) Waveform inversion of volcano-seismic signals assuming possible source geometries. *Geophys Res Lett* 32:L12302. <https://doi.org/10.1029/2005GL022666>
- Nakano M, Kumagai H, Kumazawa M, Yamaoka K, Chouet BA (1998) The excitation and characteristic frequency of the long-period volcanic event: an approach based on an inhomogeneous autoregressive model of a linear dynamic system. *J Geophys Res* 103:10031–10046
- Nakano M, Kumagai H, Chouet BA (2003) Source mechanism of long-period events at Kusatsu-Shirane volcano, Japan, inferred from waveform inversion of the effective excitation functions. *J Volcanol Geotherm Res* 122:149–164
- Nakano M, Kumagai H, Chouet BA, Dawson PB (2007) Waveform inversion of volcano-seismic signals for an extended source. *J Geophys Res* 112:B02306. <https://doi.org/10.1029/2006JB004490>
- Neuberg JW (2000) Characteristics and causes of shallow seismicity in andesite volcanoes. *Philos Trans R Soc Lond A* 358:1533–1546
- Neuberg JW, Luckett R, Ripepe M, Braun T (1994) Highlights from a seismic broadband array on Stromboli Volcano. *Geophys Res Lett* 21:749–752
- Neuberg JW, Luckett R, Baptie B, Olsen K (2000) Models of tremor and low-frequency earthquake swarms on Montserrat. *J Volcanol Geotherm Res* 101:83–104
- Neuberg JW, Tuffen H, Collier L, Green D, Powell T, Dingwell D (2006) The trigger mechanism of low frequency earthquakes on Montserrat. *J Volcanol Geotherm Res* 153:37–50

- Nishimura T (2004) Pressure recovery in magma due to bubble growth. *Geophys Res Lett* 31:L12613. <https://doi.org/10.1029/2004GL019810>
- Nishimura T, Nakamichi H, Tanaka S, Sato M, Kobayashi T, Ueki S, Hamaguchi H, Ohtake M, Sato H (2000) Source process of very long period seismic events associated with the 1998 activity of Iwate volcano, Northeastern Japan. *J Geophys Res* 105:19135–19417
- Nishimura T, Ueki S, Yamawaki T, Tanaka S, Hashino H, Sato M, Nakamichi H, Hamaguchi H (2002) Broadband seismic signals associated with the 2000 volcanic unrest of mount Bandai, northeastern Japan. *J Volcanol Geotherm Res* 119:51–59
- O'Brien GS, Bean CJ (2004) A 3D discrete numerical elastic lattice method for seismic wave propagation in heterogeneous media with topography. *Geophys Res Lett* 31:L14608. <https://doi.org/10.1029/2004GL020069>
- Ohminato T (2006) Characteristics and source modeling of broadband seismic signals associated with the hydrothermal system at Satsuma-Iwojima volcano, Japan. *J Volcanol Geotherm Res* 158:467–490
- Ohminato T, Chouet BA (1997) A free-surface boundary condition for including 3D topography in the finite difference method. *Bull Seism Soc Am* 87:494–515
- Ohminato T, Chouet BA, Dawson PB, Kedar S (1998) Waveform inversion of very long period impulsive signals associated with magmatic injection beneath Kilauea volcano, Hawaii. *J Geophys Res* 103:23839–23862
- Ohminato T, Takeo M, Kumagai H, Yamashina T, Oikawa J, Koyama E, Tsuji H, Urabe T (2006) Vulcanian eruptions with dominant single force components observed during the Asama 2004 volcanic activity in Japan. *Earth Planets Space* 58:583–593
- Ripperger J, Igel H, Wasserman J (2003) Seismic wave simulation in the presence of real volcano topography. *J Volcanol Geotherm Res* 128:31–44
- Rowe CA, Aster RC, Kyle PR, Schlue JW, Dibble RR (1998) Broadband recording of Strombolian explosions and associated very-long-period seismic signals on mount Erebus volcano, Ross Island, Antarctica. *Geophys Res Lett* 25:2297–2300
- Rubin AM, Gillard D (1998) Dike-induced seismicity: theoretical considerations. *J Geophys Res* 103:10017–10030
- Rubin AM, Gillard D, Got JL (1998) A reinterpretation of seismicity associated with the January 1983 dike intrusion at Kilauea volcano, Hawaii. *J Geophys Res* 103:10003–10015
- Saccorotti G, Del Pezzo E (2000) A probabilistic approach to the inversion of data from a seismic array and its application to volcanic signals. *Geophys J Int* 143:249–261
- Saccorotti G, Chouet BA, Dawson PB (2001) Wavefield properties of a shallow long-period event and tremor at Kilauea volcano, Hawaii. *J Volcanol Geotherm Res* 109:163–189
- Sakuraba A, Oikawa J, Imanishi Y (2002) Free oscillations of a fluid sphere in an infinite elastic medium and long-period volcanic earthquakes. *Earth Planets Space* 54:91–106
- Stump BW, Johnson LR (1977) The determination of source properties by the linear inversion of seismograms. *Bull Seism Soc Am* 67:1489–1502
- Sturton S, Neuberg JW (2006) The effects of conduit length and acoustic velocity on conduit resonance: implications for low-frequency events. *J Volcanol Geotherm Res* 151:319–339
- Takei Y, Kumazawa M (1994) Why have the single force and torque been excluded from seismic source models? *Geophys J Int* 118:20–30
- Takei Y, Kumazawa M (1995) Phenomenological representation and kinematics of general seismic sources including the seismic vector modes. *Geophys J Int* 121:641–662
- Takeo M, Yamasato H, Furuya I, Seino M (1990) Analysis of long-period seismic waves excited by the November 1987 eruption of Izu-Oshima volcano. *J Geophys Res* 95:19377–19393
- Takeuchi H, Saito M (1972) Seismic surface waves. In: Bolt BA (ed) *Methods in computational physics*, vol 11. Academic, New York, pp 217–295
- Tameguri T, Iguchi M, Ishihara K (2002) Mechanism of explosive eruptions from moment tensor analyses of explosion earthquakes at Sakurajima volcano, Japan. *Bull Volcanol Soc Jpn* 47:197–216
- Toda S, Stein R, Sagiya T (2002) Evidence from the AD 2000 Izu islands earthquake swarm that stressing rate governs seismicity. *Nature* 419:58–61
- Tolstoy I (1954) Dispersive properties of a fluid layer overlying a semi-infinite elastic solid. *Bull Seism Soc Am* 44:493–512
- Tuffen H, Dingwell D (2005) Fault textures in volcanic conduits: evidence for seismic trigger mechanisms during silicic eruptions. *Bull Volcanol* 67:370–387
- Uhira K, Takeo M (1994) The source of explosive eruptions of Sakurajima volcano, Japan. *J Geophys Res* 99:17775–17789
- Ulrych TJ, Clayton RW (1976) Time series modeling and maximum entropy. *Phys Earth Planet Inter* 12:188–200
- Ulrych TJ, Sacchi MD (1995) Sompi, Pisarenko and the extended information criterion. *Geophys J Int* 122:719–724
- Virieux J (1986) P-SV wave propagation in heterogeneous media: velocity-stress finite-difference method. *Geophysics* 51:889–901
- Wessel P, Smith WHF (1998) New improved version of generic mapping tools released. *Eos Trans AGU* 122:149–164
- Yamamoto M (2005) Volcanic fluid system inferred from broadband seismic signals. Ph D thesis. In: University of Tokyo

- Yamamoto M, Kawakatsu H, Kaneshima S, Mori T, Tsutsui T, Sudo Y, Morita Y (1999) Detection of a crack-like conduit beneath active crater at Aso volcano, Japan. *Geophys Res Lett* 26:3677–3680
- Yamamoto M, Kawakatsu H, Yomogida K, Koyama J (2002) Long-period (12 sec) volcanic tremor observed at Usu 2000 eruption: seismological detection of a deep magma plumbing system. *Geophys Res Lett* 29:1329. <https://doi.org/10.1029/2001GL013996>
- Yamamura K, Kawakatsu H (1998) Normal-mode solutions for radiation boundary conditions with an impedance contrast. *Geophys J Int* 134:849–855

Books and Reviews

- Dahlen FA, Tromp J (1998) *Theoretical global seismology*. Princeton University Press, Princeton
- Kay SM, Marple SL (1981) Spectrum analysis – a modern perspective. *Proc IEEE* 69:1380–1419

Volcanoes, Non-linear Processes in

Bernard Chouet
US Geological Survey, Menlo Park, USA

Article Outline

Glossary
Definition of the Subject
Introduction
Description of Seismic Sources in Volcanoes
Sources of Long-Period Seismicity
Source Processes of Very-Long-Period Signals
Future Directions
Bibliography

Glossary

Bubbly liquid Liquid-gas mixture in which the gas phase is distributed as discrete bubbles dispersed in a continuous liquid phase.

Choked flow Flow condition where the transonic flow of a compressible fluid through a constriction becomes choked. As the fluid flows through a narrowing channel, it accelerates at a rate that depends on the ratio of channel inlet pressure to channel outlet pressure. At a critical pressure ratio, which is a function of the fractional change of cross-sectional area, the speed of the fluid accelerating through the nozzle-like constriction reaches a maximum value equal to its sound speed. This flow condition is termed choked flow.

Crack wave A dispersive wave generated by fluid-solid interaction in a fluid-filled crack embedded in an elastic solid. The crack-wave speed is always smaller than the acoustic speed of the fluid.

Diffuse interface Thin mixing layer constituting the interface between two immiscible fluids. Interface dynamics is governed by molecular

forces and is described by the mixing energy of the two fluid components.

Gas slug A gas bubble whose diameter is approximately the diameter of the pipe in which the slug is flowing. In a slug ascending a vertical pipe, the nose of the bubble has a characteristic spherical cap and the gas in the rising bubble is separated from the pipe wall by a falling film of liquid.

Long-period (LP) event A seismic event originating under volcanoes with emergent onset of *P* waves, no distinct *S* waves, and a harmonic signature with periods in the range 0.2–2 s as compared to a tectonic earthquake of the same magnitude. LP events are attributed to the involvement of fluid such as magma and/or water in the source process (see also VLP event).

Magma Molten rock containing variable amounts of dissolved gases (principally water vapor, carbon dioxide, and sulfur dioxide), crystals (principally silicates and oxides), and, occasionally, preexisting solid rock fragments.

Moment tensor A symmetric second-order tensor that completely characterizes an internal seismic point source. For an extended source, it represents a pointsource approximation and can be quantified from an analysis of seismic waves whose wavelengths are much longer than the source dimensions.

Nonlinear process Process involving physical variables governed by nonlinear equations that reflect the fundamental micro-scale dynamics of the system. Nonlinear phenomena can arise from a number of different sources, including convective acceleration in fluid dynamics, constitutive relations, boundary conditions representing gas-liquid interfaces, nonlinear body forces, and geometric nonlinearities arising from large-scale deformation.

Phase-field method Method treating the interface between two immiscible fluids as a diffuse thin layer; also known as the diffuse interface method. In this approach, the physical and chemical properties of the interfacial layer are

defined by a phase-field variable ϕ , whose dynamics are expressed by the Cahn–Hilliard convection-diffusion equation. Use of the phase-field variable avoids the necessity of tracking the interface and yields the correct interfacial tension from the free energy stored in the mixing layer.

Very-long-period (VLP) event A seismic event originating under volcanoes with typical periods in the range 2–100 s. VLP events are attributed to the involvement of fluid such as magma and/or water in the source process. Together with LP events, VLP events represent a continuum of fluid oscillations, including acoustic and inertial effects resulting from perturbations in the flow of fluid through conduits.

Volatile A chemical compound that is dissolved in magma at high pressure and exsolves from the melt and appears as a low-density gas at low pressure. The most common volatiles in magma are water, carbon dioxide, and sulfur dioxide.

Waveform inversion Given an assumed model of the wave-propagation medium, a procedure for determining the source mechanism and source location of a seismic event based on matching observed waveforms with synthetics calculated with the model.

Definition of the Subject

Magma transport is fundamentally episodic in character as a result of the inherent instability of magmatic systems at all time scales. This episodicity is reflected in seismic activity, which originates in dynamic interactions between gas, liquid and solid along magma transport paths involving complex geometries. The geometrical complexity plays a central role in controlling flow disturbances and also providing specific sites where pressure and momentum changes in the fluid are effectively coupled to the Earth.

Recent technological developments and improvements in the seismological instrumentation of volcanoes now allow the surface effects of subterranean volcanic processes to be imaged in unprecedented detail. Through detailed analyzes

of the seismic wavefields radiated by volcanic activity, it has become possible for volcano seismologists to make direct measurements of the volcanic conduit response to flow processes, thus opening the way for detailed modeling of such processes.

However, unlike the description of seismic waves, which is based on the linear equations of elastodynamics, the description of the flow processes underlying the observed seismic source mechanisms is governed by the nonlinear equations of fluid dynamics. In the classic Navier–Stokes equations of fluid mechanics, the nonlinearity resides in the convective acceleration terms in the equations of conservation of momentum. In volcanic fluids, further complexity arises from the strong nonlinear dependence of magma rheology on temperature, pressure, and water and crystal content, and nonlinear characteristics of associated processes underlying the physico-chemical evolution of liquid-gas mixtures constituting magma. An example of nonlinearity in two-phase fluid mixtures are the changing boundary conditions associated with the internal surfaces separating the phases, a situation commonly encountered during the spinodal decomposition of binary fluids (Jones 2002). Spinodal decomposition occurs when a mixture of two species *A* and *B* of fluid forming a homogeneous mixture at high temperature undergoes spontaneous demixing following a temperature drop (quench) below some critical temperature. As the temperature drops below the critical temperature, the initially homogeneous mixture becomes locally unstable, and a period of interdiffusion is initiated, which leads to the formation of patches of A-rich and B-rich fluids separated by sharp interfaces. In the late stages of demixing following a deep quench, the interfacial thickness reaches a molecular scale and the interfacial tension approaches an equilibrium surface tension. As it approaches local equilibrium, the mixture continues to evolve toward a minimization of energy by reducing its interfacial area, and stresses arising from changes in the local interfacial curvature drive fluid motion. A smooth evolution of the interfacial structure ensues, occasionally interrupted by local coalescence, breakup and reconnection, until final equilibrium is

reached between the two bulk domains. Similar processes underlie the physics that governs the vesiculation, fragmentation, and collapse of bubble-rich suspensions to form separate melt and vapor in response to decreasing pressure along the ascent path of magma.

All of the above factors contribute to a diversity of oscillatory processes that intensifies as magma rheology becomes more complex; increasingly diverse behavior also characterizes hydrothermal fluids. Heat transfer to the hydrothermal system from magmatic gases streaming through networks of fractures pervading the rock matrix may induce boiling in ground water, resulting in the formation of steam bubbles and attendant bubble oscillations, thereby generating sustained tremor of hydrothermal origin.

Refined understanding of magma and hydrothermal transport dynamics therefore requires multidisciplinary research involving detailed field measurements, laboratory experiments, and numerical modeling. The comprehensive breadth of seismology must be used to understand and interpret the wide variety of seismic signals encountered in various volcanic processes. This effort must then be complemented by laboratory experiments and numerical modeling to better understand the complex flow dynamics at the origin of the source mechanisms imaged from seismic data. Such research is fundamental to monitoring and interpreting the subsurface migration of magma that often leads to eruptions, and thus enhances our ability to forecast hazardous volcanic activity.

Introduction

Degassing is the main driving force behind most volcanic phenomena. The separation of vapor and melt phases leads to the formation of bubbles, the presence of which decreases magma density, enhances magma buoyancy and propels magma ascent (Wilson and Head 1981). Bubbles in magma also naturally increase magma compressibility, resulting in a low acoustic speed of the bubbly mixture compared to that of the liquid phase (Chouet 1996b; Collier et al. 2006;

Commander and Prosperetti 1989; Ichihara and Kameda 2004). At shallow depths, the high compressibility of the bubbles may further contribute to the generation of pressure oscillations in bubble clouds (Chouet 1996b; Yoon et al. 1991). Bubble coalescence and fragmentation can also contribute to pressure disturbances (Lane et al. 2001). In hydrothermal fluids, the collapse of small vapor bubbles can be the source of pressure pulses and sustained tremor (Kedar et al. 1996; Kieffer 1977).

Bubble dynamics play a key role in the transport of magmatic and hydrothermal fluids, not only as sources of acoustic energy, but also in providing a sharp contrast in velocity between the fluid and encasing solid, which facilitates the entrapment of acoustic energy at the source. This aspect of the source often manifests itself in the form of long-lasting, long-period harmonic oscillations produced by sustained resonance at the source.

Other types of gaseous fluid mixtures also favor source resonance. For example, gases laden with solid particles, or gases mixed with liquid droplets, may produce velocity contrasts that are similar to, or stronger, than those associated with bubbly liquids. Indeed, dusty gases made of micron-sized particles, or misty gases made of micron-sized droplets, can sustain resonance at the source over durations that far exceed those achieved with bubbly fluids (Kumagai and Chouet 2000).

Large gas slugs bursting at the liquid surface can act as active sources of acoustic and seismic radiation (James et al. 2004; Ripepe et al. 1996; Vergnolle et al. 1996). Gas slugs traversing discontinuities in conduit geometry can also cause transient liquid motions inducing pressure and momentum changes (James et al. 2006), which radiate elastic waves via their coupling to the conduit walls (Chouet et al. 2003, 2008; Ohminato et al. 1998). In magmas where a high fluid viscosity impedes flow, the diffusion of volatiles from a supersaturated melt may result in gradual pressurization of the melt and attendant deformation of the encasing rock matrix (Chouet et al. 2005, 2006; Nishimura 2004; Shimomura et al. 2006). The fluid motions resulting from

unsteady slug dynamics, large degassing bursts, and diffusion-dominated processes typically produce signals with characteristic periods longer than those commonly associated with acoustic resonance.

Oscillatory processes are thus ubiquitous during magma flow and are a natural expression of the release of thermo-chemical and gravitational energy from volcanic fluids. The radiated elastic energy reflects both active and passive processes at the source, and features a large variety of signals over a wide range of periods.

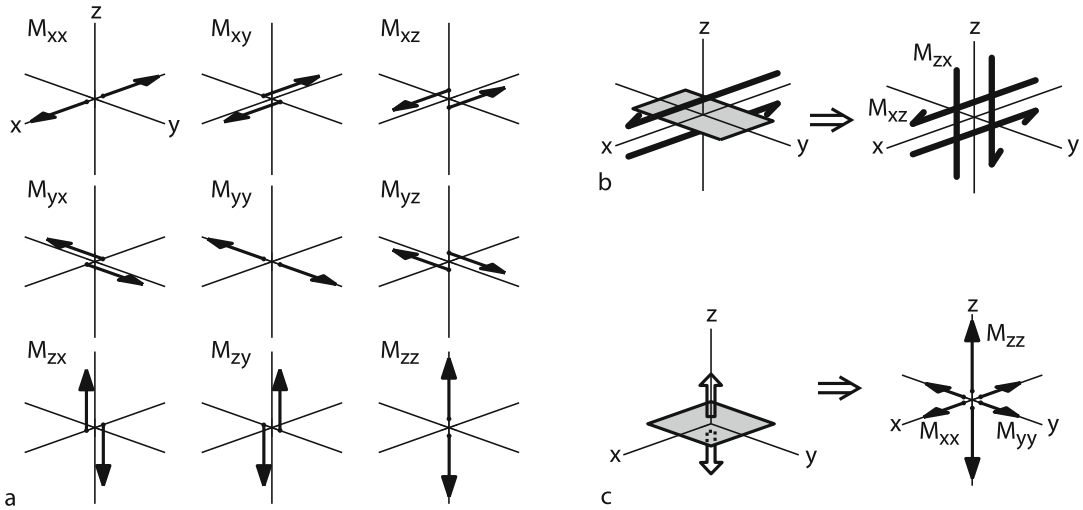
Seismic signals originating in the dynamics of magmatic and hydrothermal fluids typically include Long-Period (LP) events and tremor (Chouet 1996a, b). LP events resemble small tectonic earthquakes in duration but differ in their characteristic frequency range and harmonic signature. Tremor is characterized by a harmonic signal of sustained amplitude lasting from minutes to days, and sometimes for months or even longer. In many instances, LP events and tremor are found to have essentially the same temporal and spectral components, suggesting that a common source process, differing only in duration, underlies these two types of events. Accordingly, LP events and tremor are often grouped under the common appellation LP seismicity. The periods in which LP seismicity is observed typically range from 0.2 to 2 s, and the characteristic oscillations of LP signals are commonly viewed as a result of acoustic resonance in a fluid-filled cavity. Slower processes involving inertial forces associated with unsteady mass transport may radiate seismic signals extending to periods much longer than 2 s. These types of signals commonly fall under the appellation of Very-Long-Period (VLP) seismicity. Collectively, LP and VLP seismicities provide a comprehensive view of mass transport dynamics under a volcano.

The present chapter offers a brief review of the state of the art in volcano seismology and addresses basic issues in the quantitative interpretation of processes operative in active volcanic systems. The chapter starts with an introduction of the seismic methodology used to quantify the source of volcano seismicity, with emphasis on sources originating in the dynamics of volcanic

fluids. A review of some of the representative source mechanisms of LP and VLP signals and of their implications for volcanic processes follows in sections “Sources of Long-Period Seismicity” and “Source Processes of Very-Long-Period Signals”. A final section describes a meso-scale computational approach for simulating two-phase flows of complex magmatic fluids. This method, called the phase-field method, grew out of the original work of (Cahn and Hilliard 1958), and relies on an energy-based variational formalism to treat interfacial dynamics and complex rheology in a unified framework. The phase-field method has been successfully applied to the modeling of complex micro-structured two-phase fluids in the field of engineering materials and appears well adapted to the numerical simulation of nonlinear volcanic processes.

Description of Seismic Sources in Volcanoes

A general kinematic description of seismic sources in volcanoes is commonly based on a moment-tensor and single-force representation of the source (Aki and Richards 1980) (Fig. 1a). The seismic-moment tensor consists of nine force couples, with each corresponding to one set of opposing forces (dipoles or shear couples). This symmetric second-order tensor allows a description of any generally oriented discontinuity in the Earth in terms of equivalent body forces. For example, slip on a fault can be represented by an equivalent force system involving a superposition of two force couples of equal magnitudes – a double couple (Fig. 1b). Similarly, a tensile crack has an equivalent force system made of three vector dipoles, with dipole magnitudes with ratios $1:1:(\lambda + 2\mu)/\lambda$, in which λ and μ are the Lamé coefficients of the rock matrix (Chouet 1996b), and where the dominant dipole is oriented normal to the crack plane (Fig. 1c). Injection of fluid into the crack will cause the crack to expand and act as a seismic source. In general, magma movement between adjacent segments of conduit can be represented through a combination of volumetric sources of this type. Because



Volcanoes, Non-linear Processes in, Fig. 1 (a) The nine possible couples corresponding to the moment-tensor components describing the equivalent force system for a seismic source in the Earth. (b) Slip on a fault can be described by a superposition of two force couples, in which each force couple is represented by a pair of forces

offset in the direction normal to the force. Sources involving slip on a fault thus have an equivalent force system in the form of a double couple composed of four forces. (c) A tensile crack has a representation in the form of three vector dipoles, in which each dipole consists of a pair of forces offset in the direction of the force

mass-advection processes can also generate forces on the Earth, a complete description of volcanic sources commonly requires the consideration of single forces in addition to the volumetric source components expressed in the moment tensor. For example, a volcanic eruption can induce a force system that includes a contraction of the conduit/reservoir system in response to the ejection of fluid, and a reaction force from the eruption jet (Kanamori et al. 1984) (Fig. 2). Some volcanic processes can be described by a single-force mechanism only. An example is the single-force mechanism attributed to the massive landslide observed at the start of the 1980 eruption of Mount St. Helens (Kanamori and Given 1982; Kawakatsu 1989). A single-force source model was also proposed by Uhira et al. (1994) to explain the mechanism of dome collapses at Unzen Volcano, Japan. In general terms, a single force can be generated by an exchange of linear momentum between the source and the rest of the Earth (Takei and Kumazawa 1994). Thus, a single force on the Earth may result from an acceleration of the center of mass of the source in the direction opposite to the force, or deceleration of the center of mass of the source in the same direction as the force (Takei and Kumazawa 1994).

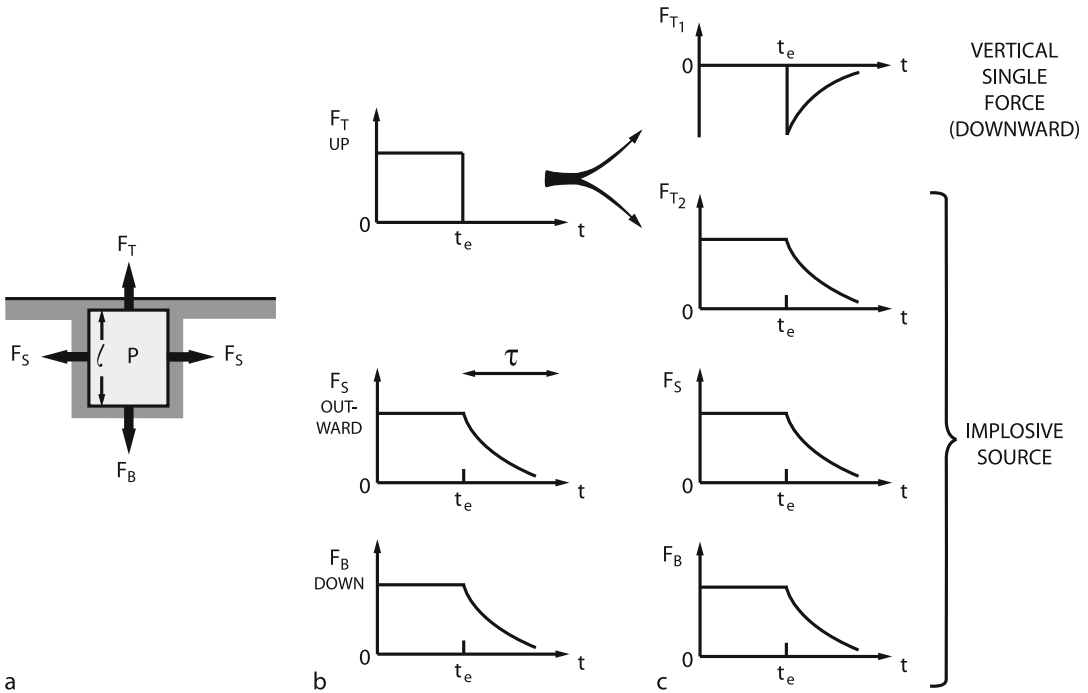
Waveform Inversion

Waveform inversions solving for the amplitudes and time histories of the moment-tensor and single-force components of the source have become the primary tool to identify and understand the mechanisms of generation of LP and VLP signals recorded by broadband seismometers. When the wavelengths of observed seismic waves are much longer than the spatial extent of the source, the source may be approximated by a point source, and the force system represented by the moment-tensor and single-force components is localized at this point. The waveform inversion procedure requires a calculation of the impulse responses (Green's functions) of the medium to this force system for each receiver component. The displacement field generated by a seismic source is described by the representation theorem which, for a point source, may be written as e.g., Chouet (1996b)

$$u_n(t) = F_p(t) * G_{np}(t) + M_{pq}(t) * G_{np,q}(t),$$

$$p, q = x, y, z, \quad (1)$$

where $u_n(t)$ is the n -component of seismic displacement at a receiver at time t , $F_p(t)$ is the time history of the force applied in the p -direction,



Volcanoes, Non-linear Processes in, Fig. 2 Force system equivalent to a volcanic eruption (after Kanamori et al. 1984). (a) Pressurized cavity model for a volcanic eruption. A shallow vertically oriented cylindrical cavity initially sealed at the top by a lid contains a pressurized inviscid fluid which exerts an upward vertical force F_T on the lid, a horizontal outward force F_S on the sidewall, and a vertical downward force F_B on the bottom of the cylinder. (b) Time histories of forces acting on the top, side, and bottom cavity walls of the cavity. The eruption is simulated by the sudden removal of the lid at time $t = t_e$, at which point the force F_T vanishes instantaneously and the fluid pressure in the cylinder starts to decrease with a characteristic time constant τ fixed by the mass flux of the eruption,

i.e., $\tau \sim \ell/v$, where ℓ is the length of the cylinder and v is the mean fluid velocity inside the cylinder. Since the forces F_S and F_B are both proportional to pressure, they decrease with the same time constant τ . (c) Decomposition of the force system to a downward single force and implosive source. The force F_T in b is decomposed into a vertical downward component F_{T1} and vertical upward component F_{T2} in such a way that F_{T1} has the same time history as F_S and F_B . As a result, the three forces F_{T2} , F_S , and F_B form an implosive source so that the eruption mechanism is represented by the superposition of a downward vertical force (the reaction force of the volcanic jet) with this volumetric implosion

$M_{pq}(t)$ is the time history of the pq -component of the moment tensor, and $G_{np}(t)$ is the Green tensor which relates the n -component of displacement at the receiver position with the p -component of impulsive force at the source position. The notation, q indicates spatial differentiation with respect to the q -coordinate and the symbol $*$ denotes convolution. Summation over repeated indices is implied.

A common approach to the quantification of the seismic source mechanism relies on a discretized representation of the volcanic edifice based on a digital elevation model. Using this model, Green's functions are then calculated by

the finite difference method (Ohminato and Chouet 1997). Once the Green's functions are known, the nine independent mechanisms in Eq. (1) are retrieved through least-squares inversion expressing the standard linear problem $\mathbf{d} = \mathbf{G}\mathbf{m}$ in this equation (Chouet et al. 2003).

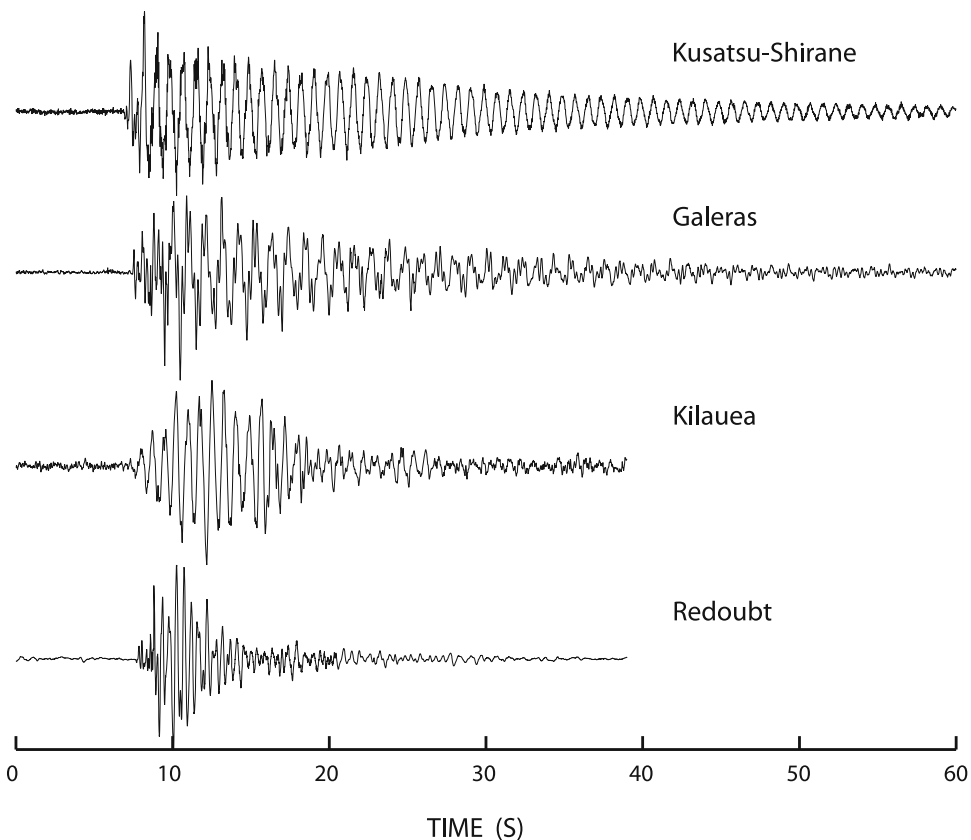
Sources of Long-Period Seismicity

Long-Period (LP) seismicity, including individual LP events and tremor, is commonly observed in relation to magmatic and hydrothermal activities in volcanic areas and is recognized as a

precursory phenomenon for eruptive activity (e.g., Chouet 1996a). The waveform of the LP event is characterized by simple decaying harmonic oscillations except for a brief interval at the event onset (Fig. 3). This signature is commonly interpreted as oscillations of a fluid-filled resonator in response to a time-localized excitation. By the same token, tremor may be viewed as oscillations of the same resonator in response to a sustained excitation. LP events are particularly useful in the quantification of magmatic and hydrothermal processes because the properties of the resonator system at the source of this event can be inferred from the properties of the decaying oscillations in the tail of the seismogram. The damped oscillations in the LP coda are quantified by two parameters, T , and Q , where T is the period of the dominant mode of

oscillation, and Q is the quality factor of the oscillatory system representing the combined effects of intrinsic and radiation losses.

Interpretations of the oscillating characteristics of LP sources have mostly relied on a model of fluid-driven crack (Chouet 1986, 1988, 1992). This model, which has the most natural geometry that satisfies mass-transport conditions at depth beneath a volcano, is supported by results from inversions of LP waveforms recorded at several volcanoes (Kumagai et al. 2002b, 2005; Nakano et al. 2003; Waite et al. 2008). A simplified two-dimensional model of fluid-driven crack was originally proposed by Aki et al. (1977). Although this model considered both the driving excitation and geometry appropriate for transport, the fluid inside the crack was only treated as a passive cushion that did not support the acoustic



Volcanoes, Non-linear Processes in, Fig. 3 Typical signatures of long-period events observed at Kusatsu-Shirane, Galeras, Kilauea, and Redoubt Volcanoes. The

signatures are all characterized by a harmonic coda following a signal onset enriched in higher-frequencies

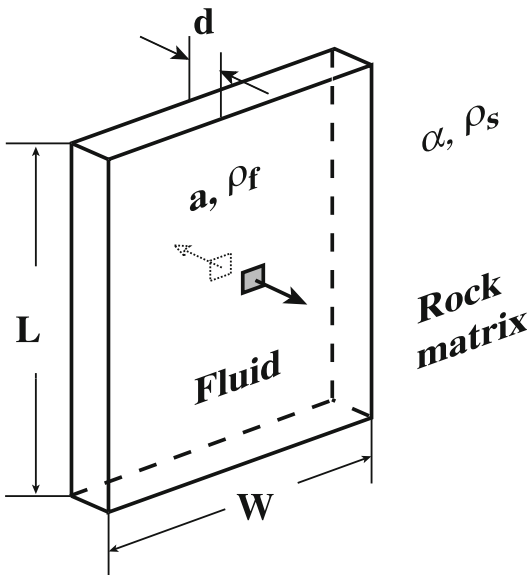
propagation of the pressure disturbance caused by the motion of the crack wall. An extension of this model including active fluid participation was proposed by Chouet and Julian (1985), who considered a simultaneous solution of the elastodynamics and fluid dynamics for a two-dimensional crack. This model was further extended to three dimensions by Chouet (1986) and was extensively studied by Chouet (1988, 1992).

Chouet's three-dimensional model consists of a single isolated rectangular crack embedded in an infinite elastic body and assumes zero mass transfer in and out of the crack (Fig. 4). Crack resonance is excited by a pressure transient applied symmetrically on both walls over a small patch of crack wall. In this model, the crack aperture is assumed to be much smaller than the seismic wavelengths of interest and the motion of the fluid inside the crack is treated as two-dimensional in-plane motion. The fluid dynamics

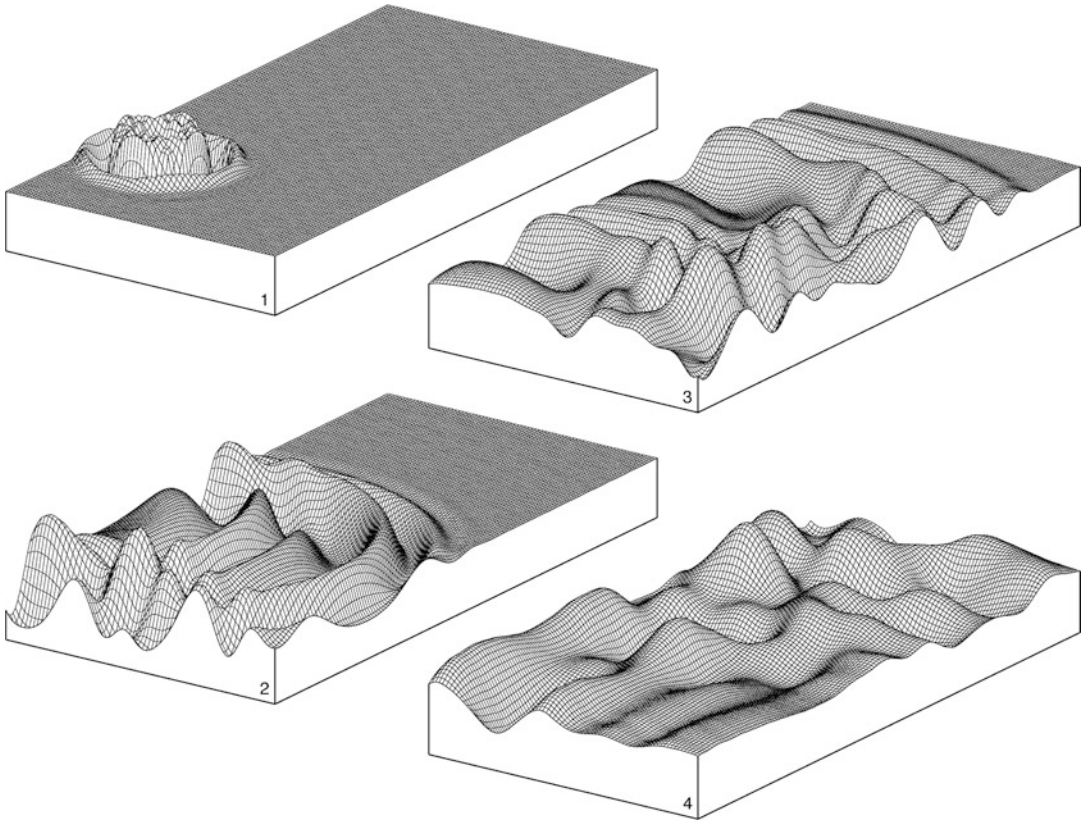
are represented by the conservation of mass and momentum for the two in-plane components of fluid velocity averaged over the crack aperture, and convective terms in the momentum equations are neglected assuming their contributions are negligibly small compared to the time derivatives of pressure and flow velocity. A solution of the coupled equations of fluid dynamics and elastodynamics is then obtained by using a finite-difference approach. As formulated, the model accounts for the radiation loss only; intrinsic losses due to dissipation mechanisms within the fluid are treated separately (Kumagai and Chouet 2000).

Figure 5 shows snapshots of the normal component of particle velocity on the crack wall obtained at four different times following the onset of crack resonance. In this particular example, the crack excitation is triggered by a step in pressure applied at a point located a quarter of the crack width from the left edge of the crack and a quarter of the crack length from the front edge of the crack. The first snapshot at the upper left shows the propagation of the elastic disturbance on the crack surface shortly after the onset of the pressure transient. The second snapshot at the lower left shows the motion of the crack wall following the first reflections from the front, left and right edges of the crack. The third snapshot at the upper right represents the wall motion immediately preceding the reflection from the back edge of the crack, and the fourth snapshot at the lower right depicts the shape of the lateral and longitudinal modes of crack resonance, which by then are well established.

The excitation of modes in the crack model depends on the position of the pressure transient, the extent of crack surface affected by the transient, the temporal characteristics of the transient, and crack stiffness $C = (b/\mu)(L/d)$, where b is the bulk modulus of the fluid, μ is the rigidity of the rock, L is the crack length, and d is the crack aperture (Aki et al. 1977; Chouet 1986). An example of synthetic seismo-gram representing the vertical ground velocity response in the near field of a fluid-filled crack excited into resonance by a step transient in pressure is shown in Fig. 6. The synthetics display sustained oscillations that bear

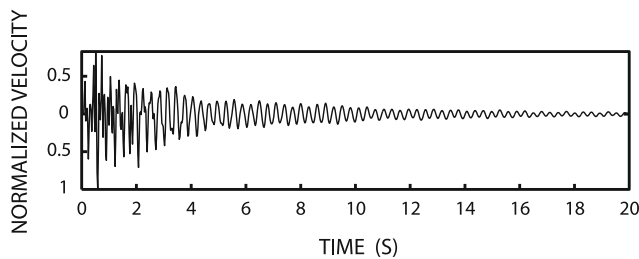


Volcanoes, Non-linear Processes in, Fig. 4 Geometry of the fluid-filled crack model of Chouet (1986). The crack has length L , width W , and aperture d , and contains a fluid with sound speed a and density ρ_f . The crack is embedded in an elastic solid with compressional wave velocity α and density ρ_s . Excitation of the crack is provided by a pressure transient applied symmetrically on both walls over the small areas such as that indicated by the gray patch and dotted small square



Volcanoes, Non-linear Processes in, Fig. 5 Snapshots of the normal component of particle velocity on the surface of a fluid-filled crack excited by a step in pressure applied over a small patch of the crack wall. The snapshots (1–4) show the propagation of the disturbance on the crack wall at four different times in the history of crack oscillation specified by the number of time steps, N , in the finite-

difference solution obtained by Chouet (1986). The time step has size $\Delta t = 0.003125L/\alpha$ (see Fig. 4 for explanation of symbols). Depicted are the shape of the wall motion at $N = 72$ (snapshot 1), $N = 312$ (snapshot 2), $N = 560$ (snapshot 3), and $N = 1200$ (snapshot 4). (See text for discussion)



Volcanoes, Non-linear Processes in, Fig. 6 Synthetic velocity waveform observed at an epicentral distance of 500 m and azimuth of 45° from the crack trace for a vertical crack (vertical extent $L = 150$ m and horizontal extent $W = 75$ m) buried at a depth of 500 m in a homogeneous

half space. The aspect ratio is $L/d = 10^4$, and the parameters of the fluid and solid are $a = 300$ m/s, $\rho_f = 2120$ kg/m³, and $\alpha = 4500$ m/s, $\rho_s = 2650$ kg/m³. (Reproduced from Kumagai and Chouet 2000)

strong resemblance to observed LP waveforms (compare with Fig. 3), convincingly demonstrating the critical involvement of active fluid participation in the source process of volcanic signals.

Chouet's studies demonstrated the existence of a slow wave propagating along the crack wall, which he named the "crack wave". The asymptotic behavior of this wave was investigated by Ferrazini and Aki (1987) in an analytical study of normal modes trapped in a liquid layer sandwiched between two elastic half spaces. The crack wave speed is slower than the sound velocity of the liquid and decreases with increasing crack stiffness and increasing wavelength; this result reflects the elasticity of the crack wall, whose effect is to decrease the effective bulk modulus of the fluid. These slow characteristics of the crack wave lead to more realistic estimates of crack dimensions compared to estimates based on the sound speed of a fluid embedded in a resonator with perfectly rigid walls. Source dimensions estimated from LP data based on this model typically range from tens to several hundred meters (Kumagai et al. 2002a, 2005; Saccorotti et al. 2001). Detailed analyses of the dependence of crack resonance on fluid composition carried out by Kumagai and Chouet (2000), and systematic investigations of LP signatures based on their results, suggest that dusty gases or bubbly basalt are common fluids in LP events of magmatic origin (Chouet 1996a; Gil Cruz and Chouet 1997; Kumagai and Chouet 1999), and that wet gases, steam and bubbly water are typically representative of the source of LP events of hydrothermal origin (Kumagai et al. 2002a; Saccorotti et al. 2001).

The model of Chouet (1986) does not address the excitation mechanism of LP events or tremor. Rather, the spatiotemporal properties of the pressure transient triggering the crack resonance are preset as kinematic conditions in the model. The usefulness of this model is thus restricted to a quantification of the crack resonance and properties of the fluids at the source of LP events. For a better perspective of the excitation mechanism of LP seismicity, observations of the source over a wider band of frequencies are necessary, along with an adequate sampling of the evolutionary

characteristics of volcanic seismicity. Observations carried out in different volcanic settings point to a wide variety of LP excitation mechanisms originating in magmatic-hydrothermal interactions, as well as magmatic instabilities.

Magmatic-Hydrothermal Interactions

In hydrothermal systems, LP seismicity can be induced by surges in the heat transfer from an underlying magma body. Kusatsu-Shirane Volcano in central Japan provides a prime example of the character of such interactions. Three crater lakes (Yugama, Mizugama, and Karagama) occupy the summit of the volcano along with numerous hot springs, pointing to an active magmatic system and enhanced circulation of hydrothermal fluids. Geochemical studies provide further support for hydrothermal activity resulting from the interaction of hot volcanic gases with groundwater (Hirabayashi 1999; Ohba et al. 2000). LP events marked by nearly monochromatic oscillatory signatures have been frequently observed at this volcano (see Fig. 3) (Fujita et al. 1995; Hamada et al. 1976), and temporal variations in the decay characteristics of these events have also been documented (Nakano et al. 1998). Kumagai et al. (2002a) performed a systematic study of the temporal characteristics of LP events observed at Kusatsu-Shirane over a period from August 1992 through January 1993, and Kumagai et al. (2002b) carried out waveform inversions of a typical event during this time interval. Results from these studies point to the repeated excitation of a fixed subhorizontal crack at depths near 200 m under the Yugama crater lake, and consistently explain temporal variations in LP signatures by the dynamic response of this crack to a magmatic heat pulse. In this interpretation, the crack contains an initially wet gas, which became gradually drier with time, suggesting a "drying out" of the hydrothermal system in response to the heat pulse. Further analyses by Nakano et al. (2003) based on waveform inversions of the effective excitation functions (the signal components remaining after removal of the resonance characteristics) of individual LP events confirm the crack mechanism imaged by Kumagai et al. (2002b) and shed further light on the source process. This

process involves a collapse and recovery of the sub-horizontal crack accompanied by an upward-directed force, and is consistent with a gradual buildup of pressure in the crack causing repeated discharges of steam from the crack. The vertical force imaged with the volumetric component of the source reflects the release of gravitational energy that occurs as a slug of steam ejected from the crack ascends toward the surface and is replaced by a downward flow of cooler water within a conduit linking the crack to the base of the Yugama crater lake. In this scenario, crack resonance is triggered by the sudden collapse of the crack induced by the venting of steam.

Similar mechanisms of interaction between magmatic and hydrothermal systems have been inferred for Kilauea, Mount St. Helens, and Redoubt Volcanoes. At Kilauea Volcano, Hawaii, Almendros et al. (2001) imaged a shallow hydrothermal system from frequency-slowness analyses of LP seismicity recorded on three small-aperture seismic arrays deployed in the summit caldera. Located within the top 500 m below the caldera floor, this reservoir is perched immediately above the shallowest segment of magma conduit identified by (Ohminato et al. 1998) from inversions of VLP signals produced during a transient in the magma flow feeding the long-lived and ongoing east rift eruption of this volcano (Heliker and Mattox 2003). Episodic deflations of the Kilauea summit during brief pauses in the flow of magma are commonly followed by rapid reinflation of the summit during the resumption of flow, and subsequent slower deflation back to an original state preceding the pause (Cervelli and Miklius 2003). Larger summit reinflation episodes associated with more energetic surges of magma are accompanied by the emission of VLP signals, production of shallow brittle failure earthquakes, and enhanced LP seismicity in the hydrothermal reservoir (Dawson et al. 2004; Ohminato et al. 1998). Detailed analyses of seismic array data by Almendros et al. (2002a) further document the persistence of low-amplitude tremor generated by shallow hydrothermal activity in response to the flux of heat from the underlying magma conduit. Additional support for this scenario also comes from analyses of the source

process of a hydrothermal LP event (Kumagai et al. 2005), which point to the resonance of a horizontal crack at depth of ~ 150 m immediately above the conduit imaged by Ohminato et al. (1998). The observed frequencies and attenuation characteristics of crack resonance are consistent with a crack filled with either bubbly water or steam (Kumagai et al. 2005).

The inferred scenario of magmatic-hydrothermal interactions at Kilauea, which is in harmony with all of the above observations, involves a chain of causes and effects that starts with a volumetric deformation of the underlying shallow segment of magma conduit induced by the surging magma. This conduit response causes brittle fractures in the surrounding rock matrix, and thus enhances the upward transport of magmatic gases through the fractured medium to shallow depths, where the hot gases heat and activate the hydrothermal system. The boiling of groundwater and attendant production of steam then raise the overall pressure of hydrothermal fluids in fractures permeating the medium, thereby leading to repeated impulsive discharges of fluid and resonant excitation of individual fractures.

The current dome-building eruption of Mount St. Helens Volcano, Washington, is marked by the extrusion of a stiff plug of dacitic magma (Dzurisin et al. 2005; Major et al. 2005) accompanied by shallow, repetitive LP events that have been dubbed “drumbeats” by scientists monitoring Mount St. Helens (Moran et al. 2007). VLP signals commonly accompany the LP events, but are generally detected only in the immediate vicinity of the crater. Using high-quality data from a temporary broadband seismic network deployed in 2004–2005, Waite et al. (2008) carried out systematic waveform inversions for both LP and VLP signals. Results from these inversions point to the perturbation of a common crack system linking the magma conduit and shallow water-saturated region of the volcano. A scenario consistent with the imaged source mechanisms involves the repeated pressurization of a sub-horizontal crack that lies beneath the growing lava dome in the southern part of the crater. Steam produced by the heating of ground water, as well as steam exsolved from the magma,

feed into the crack and pressurize it. Upon reaching a critical pressure threshold, steam is vented out of the crack through fractures leading to the surface. The drop in pressure triggers a partial collapse of the crack, which resonates for 5 to 10s of seconds. The collapse of the crack in turn induces sagging of the overlying dome and triggers a passive response in the magma conduit that is observed in the VLP band. The observed force and volumetric components of this VLP response are both consistently explained as the result of a perturbation in an otherwise smooth steady flow of magma upward through a dike and into a sill underlying the old dome. In this model, the dike represents the top of the old conduit that fed the 1980–1986 dome-building eruptions, and the sill represents a bypass below the old dome. According to this interpretation, at the start of renewed activity in September 2004, magma moving up the conduit found an easier pathway to the surface by breaking out of the conduit near the surface and pushing sub-horizontally to the south rather than intruding the older dome complex. The imaged sill represents this segment of conduit, and the hydrothermal crack represents an extension of this fracture into the shallow water-soaked region beneath the crater floor. This picture of Mount St. Helens as a steam engine is quite distinct from an earlier model in which the seismic drumbeats were assumed to represent stick-slip motion between the extruding lava and conduit walls (see below). A sustained supply of heat and fluid from the magmatic system is necessary to keep the crack pressurized and keep the drumbeats beating in the model elaborated by Waite et al. (2008). As for Kilauea, a proper interpretation of this activity has important implications for the assessment of the potential for phreatic eruptions at Mount St. Helens.

Another perspective on a different type of instability presumed to be associated with dome growth activity at Mount St. Helens was proposed by Iverson et al. (2006). Their model, originally developed before results from the detailed seismic analyses by Waite et al. (2008) became available, hypothesizes that repetitive stick-slip motion along the perimeter of the extruding solid plug may cause observable earthquakes. Although not

supported by the character of LP seismicity observed at Mount St. Helens, this model nevertheless offers an intriguing view into a possible oscillatory behavior of plug flow. Oscillations in extrusion velocity originate in the interaction of plug inertia, a variable upward force due to magma pressure, and downward force due to plug mass. Assuming that a steady ascent of compressible magma drives the upward extrusion of a solidified plug exhibiting nonlinear rate-weakening friction along its margins, Iverson et al. (2006) infer that stick-slip oscillations might be an inevitable component of such an extrusion process; however, the predicted amplitude of individual slip events, estimated at a few millimeters, suggests this mechanism may be difficult to detect. The seismic source associated with a stick-slip extrusion event may manifest itself as a combination of a near-vertical force (assuming vertical spine growth) representing the reaction force on the Earth due to the upward acceleration of mass, and double couple resulting from shear near the plug margins where viscosity is very high and magma is in the glass transition. This particular mechanism does not involve LP source resonance and is apparently not part of the dominant expression of seismicity at Mount St. Helens, but could possibly be operating at a level below the threshold of seismic detection. Such mechanism may be observable in dome-building eruptions where the extrusion rate is sufficiently large and unsteady to produce oscillations of this type.

The reawakening of Redoubt Volcano, Alaska on 13 December, 1989, after 23 years of quiescence, was heralded by a rapidly accelerating swarm of repetitive LP events that merged into sustained tremor a few hours before the eruption onset (Chouet et al. 1994; Stephens and Chouet 2001). These LP events were interpreted by Chouet et al. (1994) as the resonant excitation of a crack linking a shallow, low-pressure hydrothermal system to a deeper supercharged magma-dominated reservoir. The initiation of the swarm was attributed by these authors to the failure of a barrier separating the two reservoirs. The actual mechanism triggering the LP excitation in this model is the unsteady choking of a supersonic flow of magmatic steam driven by the pressure

gradient existing between the two reservoirs. Unsteady choking itself is the result of fluctuations in crack outlet pressure associated with the reaction of the hydrothermal system to the injection of mass and heat from below. Similarly, the emergence of sustained tremor late in the swarm is interpreted as a change in choked-flow regime related to a gradual weakening of the pressure gradient driving the flow (Meier et al. 1978). Numerical simulations of the transonic flow through a crack featuring a nozzle-like constriction were performed by Morrissey and Chouet (1997) using the Navier–Stokes equations representing a mixture of gas and suspended solid particles. In these simulations, shock waves develop immediately downstream from the nozzle-like constriction, and the flow acts as an energy transducer that converts smooth low-amplitude fluctuations of outlet flow pressure into strongly-amplified and repetitive step-like pressure transients applied to the crack wall immediately downstream from the nozzle. The magnitude of the pressure transient at the walls is fixed by the pressure of the supplying reservoir and the geometry of the nozzle aperture, presumably remaining constant as long as these conditions prevail. The temporal pattern of outlet pressure fluctuations controls the areal extent of the crack wall impacted by shock waves and, given a roughly constant shock magnitude, fixes the amplitude of the force applied to the wall by the fluid. This variable force applied at a fixed location along the crack wall is viewed as the force responsible for the scaling of amplitudes and the similarity of waveforms observed in the LP swarm at Redoubt (Chouet et al. 1994).

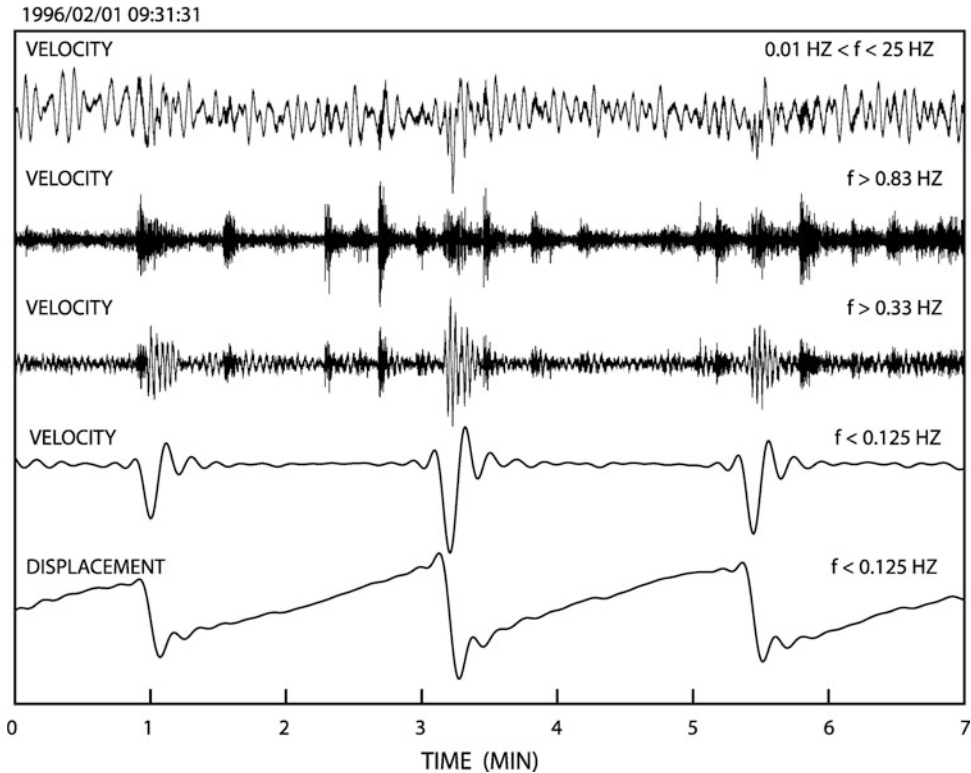
Magmatic LP Events

In low-viscosity basaltic systems, LP seismicity may be part of a broadband source process associated with the unsteady transport of magma and gases through conduits. This is well illustrated in a broadband seismic record obtained at Kilauea (Fig. 7); most interesting is the displacement record shown at bottom of figure. This record, obtained after filtering the raw broadband signal (top trace in Fig. 7) with a 0.125 Hz low-pass filter, features a repetitive sawtooth signal with

rise time of 2–3 min and drop time of 5–10 s. The observed repetitiveness of the VLP signal is a clear indication of the repeated action of a non-destructive source. Noteworthy also are the characteristic LP signatures obtained by filtering the broadband signal with a 0.33-Hz high-pass filter (third trace from the top). These LP events display a fixed dominant period of 2.5 s, whose onsets coincide with the start of the dropdown in the VLP sawtooth displacement signals. Such coincidence is strongly suggestive of a causative relationship between the VLP and LP signals.

The data in Fig. 7 are part of a seismic sequence observed during a surge in the magma flow feeding the ongoing east rift eruption of Kilauea, and the source mechanism of the VLP signals obtained by Ohmiato et al. (1998) from waveform inversions of these data points to a repeated cycle of deformation of a sub-horizontal crack located roughly 1 km below the floor of the summit caldera. Each sawtooth in the recorded displacement signal represents a slow dilation of the crack, followed by a rapid collapse of the crack. The volume change associated with individual sawtooths ranges from 1000 to 4000 m³ (Ohminato et al. 1998).

A conceptual model consistent with the seismic source mechanism imaged by Ohmiato et al. (1998) is shown in Fig. 8. This model involves the injection of a large slug of gas into a sub-horizontal crack with a narrow constricted outlet; both liquid magma and gas flow through this outlet constriction. Initially, liquid magma occupies the entire outlet cross section and gas accumulates in the crack, forming a layer of gas on top of the liquid and deforming the crack walls as the back pressure of gas increases in the crack. Eventually, when the back pressure buildup in the crack reaches a critical threshold, the gas slug deforms and rapidly squeezes past the constriction along with the liquid, triggering a rapid deflation (collapse) of the crack in the process. This sequence is then repeated with the next gas slug. In this scenario, the restraining force of the liquid on the gas acts like a self-activated viscosity-controlled valve. In the stratified flow through the nozzle, the liquid moves at a steady slow pace on the order of m/s but the gas flow itself is



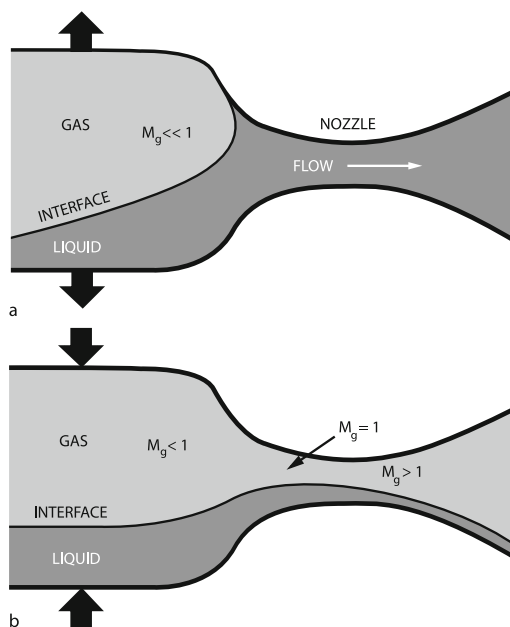
Volcanoes, Non-linear Processes in, Fig. 7 Broadband record and associated filtered signals obtained at Kilauea Volcano during a volcanic event on February 1, 1996. The broadband signal represents ground velocity and is filtered in various frequency bands to produce five records for the same 7-min time interval. The top trace shows the broadband signal ($0.01 < f < 25$ Hz), which is dominated by the oceanic wave-action microseism with dominant periods in the range 3–7 s. The second trace shows the signal obtained after application of a high-pass filter ($f > 0.83$ Hz). The result is equivalent to a typical short-period record and shows a series of events superimposed on a background of tremor. The third trace also has a high-pass filter applied

but with a lower corner frequency ($f > 0.33$ Hz); LP signals with a dominant period of about 2.5 s are enhanced in this record. The fourth trace shows the signal when a low-pass filter is applied ($f < 0.125$ Hz); a repetitive VLP signal consisting of pulses with period of about 20 s is observed. The fifth trace is the corresponding displacement record obtained with the same low-pass filter ($f < 0.125$ Hz), showing a repetitive sawtooth pattern with rise time of 2–3 min and drop time of 5–10 s. Notice that the onset of the LP signal seen above coincides with the onset of the dropdown in the VLP displacement. (Reproduced from Dawson et al. 1998)

choked in the narrow opening between the liquid/gas interface and the upper wall (Wallis 1969). The formation of a shock associated with compound choking of the flow is viewed here as a trigger of acoustic oscillations of the liquid/gas mixture, which is at the origin of the LP signature with characteristic period of 2.5 s observed in conjunction with the rapidly downgoing part of the sawtooth displacement signal shown in Fig. 7.

At Stromboli Volcano, Italy, broadband seismic signals including strong VLP and LP components are seen to accompany explosions.

Waveform inversions of the VLP signals performed by Chouet et al. (2003) show that the processes associated with eruptions involve volumetric changes marking a cycle of pressurization-depressurization-repressurization of the conduit, coupled with a single force component, both of which may be viewed as the result of a piston-like action of the magma associated with the disruption of a gas slug transiting through a suddenly enlarged conduit aperture in the steeply-inclined upper segment of conduit (Chouet et al. 2003; James et al. 2006) (see section “Slug



Volcanoes, Non-linear Processes in,

Fig. 8 Conceptual model of separated gas-liquid flow through a converging-diverging nozzle under choked conditions (see pp. 71–74 in Wallis 1969). (a) Inflation phase showing gas accumulation upstream of the nozzle, leading to pressure buildup and deformation of the crack. The gas slug is essentially stationary and the Mach number of the gas is $M_g \ll 1$. This stage coincides with the upgoing ramp in the sawtooth displacement signal at the bottom of Fig. 7. (b) Separated gas-liquid flow through the nozzle under compound choked conditions. The gas flow is choked ($M_g = 1$) in the nozzle defined by the gas-liquid interface and is supersonic ($M_g > 1$) immediately downstream of the nozzle. There exists a maximum possible gas flow rate that is fixed by the rate of liquid discharge for given upstream conditions and fixed throat geometry. In the limit, liquid fills the duct entirely and there is no gas flow. (Reproduced from Ohminato et al. 1998)

Distribution at Stromboli” for details). The LP signal in that case is attributed to the oscillatory response of the shallowest segment of the fluid-filled conduit associated with the rapid expansion and ejection of the slug (Chouet et al. 1997).

LP seismicity is also intimately linked with Vulcanian activity. The complex magma rheology, marked by a strong nonlinear dependence of viscosity on water content and temperature (Hess and Dingwell 1996; Zhang et al. 2003), finite yield strength, strain-rate dependent viscosity, and transition to brittle solid behavior at high strain rates (Tuffen et al. 2003) contributes to

widely varying oscillatory behaviors seen in Vulcanian systems. At Galeras Volcano, Colombia, LP events displaying extended quasi-monochromatic coda were found to be linked to the release of ash-laden gases from a shallow magma body through a preexisting system of cracks bisecting the dome capping the vent (Gil Cruz and Chouet 1997). Similar to Galeras, eruptive activity at Popocatepetl Volcano, Mexico, is dominated by emissions of steam, gas and ash, and by repeated formation and destruction of lava domes, all of which generate a wide variety of signals (Arciniega-Ceballos et al. 2003, 2008). LP events accompanying degassing bursts are commonly associated with VLP signals and display waveform features that remain stationary from event to event. The source of these events coincides with, or is very close to, the source of VLP signals, 1.5 km below the crater floor (Arciniega-Ceballos et al. 2008; Chouet et al. 2005). This LP activity represents an integral part of a degassing process similar to that inferred for Galeras, where the sudden expulsion of a pressurized pocket of gases induces elastic deformation of the conduit coupled with acoustic resonance of the conduit (see further discussion in section “Coupled Diffusive-Elastic Pressurization at Popocatepetl Volcano”).

LP events at Soufrière Hills, Montserrat, have been linked to the action of a repetitive, non-destructive, stationary source 1.5 km below the crater floor (Neuberg et al. 2006). The temporal relationship between LP events and tilt patterns at Soufrière Hills clearly shows that LP seismicity starts in synchronicity with the passage of tilt through a turning point marked by a maximum in time derivative of the tilt signal. Seismicity subsequently ceases as tilt goes through a second turning point (Neuberg et al. 2006). This correlation between LP events and inflation-deflation cycles is evidence of a strong relationship linking LP seismicity to magma movement and pressure conditions at depth. Based on these observations, (Neuberg et al. 2006) proposed a mechanism of LP excitation originating in the brittle failure of a highly viscous magma under high strain rate. In their model, loss of gas and heat from the magma at the conduit wall leads to the generation of a

sharp viscosity gradient at the wall. This in turn leads to a buildup of shear stress near the wall, where the flowing magma undergoing glass transition can fail in a brittle manner. The resulting brecciated zone along the conduit boundary provides a natural pathway for escaping gases. The LP events may then be viewed as an expression of acoustic resonance occurring in well-defined channels in this zone, or possibly as an expression of acoustic energy trapped in an adjacent region of conduit filled with bubbly magma (Collier et al. 2006; Ichihara and Kameda 2004).

The LP sources described so far all represent shallow sources occurring in the top 2 km below the surface. Although thoroughly documented, such sources do not represent the total expression of LP seismicity. An activity of deep (>5 km depth) LP events has also been noted in several areas of the world. In particular, episodes of deep (30–60 km) and intermediate-depth (5–15 km) LP seismicity are often observed under Kilauea (Klein et al. 1987; Koyanagi et al. 1987). Analyses of the intermittency of LP activity were carried out by Shaw and Chouet (1989) and Shaw and Chouet (1991), who viewed such tremor episodes as the relaxation oscillations of a percolation network of coupled magma-filled fractures. In their interpretation, each individual fracture in the network acts as a source of LP events, and sets of fractures give rise to multiple LP events and episodes of tremor. The aggregate nature of fracture propagation and vibration then lead to the existence of universal scaling laws that give rise to the long-recognized frequency invariance of LP events (Shaw and Chouet 1989, 1991).

Other episodic bursts have been noted at depths of 10–20 km under Mammoth Mountain, California, where LP and VLP events have been seen to occur synchronously with spasmodic bursts of small brittle failure earthquakes (Hill and Prejean 2005). A process involving the transport of a slug of CO₂-rich hydrous magmatic fluid derived from a plexus of basaltic dikes and sills has been invoked to explain such occurrences (Hill and Prejean 2005). LP events at depths near 30 km under Mount Pinatubo, Philippines, have been attributed to a basaltic intrusion thought to have triggered processes leading to the

cataclysmic eruption of June 15 (White 1996). Small LP events attributed to magmatic activity have also been detected at depths of 25–40 km beneath active volcanoes in northeastern Japan (Hasegawa et al. 1991), and at depths of 10–45 km below volcanoes of the Aleutian Arc (Power et al. 2004). In all these instances, the occurrence of deep LP events appears to be more directly related to deep magma supply dynamics than near-surface or surface activity. Because of our lack of knowledge concerning the character and dynamics of deep-seated fluid transport under volcanoes, the actual origins of these events remain enigmatic.

Summary of Inferred Excitation Mechanisms for LP Seismicity

Shallow hydrothermal LP events at Kusatsu-Shirane, Kilauea, and Mount St. Helens all share a common excitation mechanism involving the repeated pressurization of a steam-filled fracture, causing the venting of steam, collapse of the fracture and recharge, in response to heat transfer from an underlying magma body. At Mount St. Helens, where the LP source is located very close to the magma conduit, LP activity is also seen to trigger a passive response of the conduit itself. LP seismicity observed at Redoubt represents a more energetic form of magmatic-hydrothermal interaction where an unsteady choked flow of magmatic gases provides a natural source of pressure perturbation at the origin of LP events and tremor. Magmatic LP events produced during bursts of ash-laden gases associated with Vulcanian activity at Galeras and Popocatepetl involve a pumping mechanism in shallow fractures similar to that inferred for hydrothermal LP events at Kusatsu-Shirane, Kilauea, and Mount St. Helens. LP events accompanying endogenous dome growth at Soufrière Hills appear to be related to a more complex process involving the production of shear fractures in highly viscous magma at the glass transition, injection of dusty gases into these fractures, and resonance of the fracture network and/or possibly resonance of a bubble-rich magma excited by the energy release from the brittle failure. Magmatic LP events in the basaltic systems at Kilauea and Stromboli are the

result of pressure disturbances generated during the transit of large slugs of gas through conduit discontinuities.

A nonlinear excitation mechanism of LP seismicity by fluid flow has also been proposed by Julian (1994). Using a simple lumped-parameter model, Julian investigated the elastic coupling of the fluid and solid as a means to produce self-excited oscillations in a viscous incompressible liquid flowing through a channel with compliant walls. In his model, an increase in flow velocity leads to decrease of fluid pressure via the Bernoulli effect. As a result, the channel walls move inward and constrict the flow, causing an increase in fluid pressure and forcing the channel open again. The cyclic repetition of this process is the source of sustained oscillations in Julian's model. Julian demonstrated that with increasing driving pressure the model can exhibit various oscillatory behaviors resembling tremor.

As illustrated in the above examples, magmatic-hydrothermal interactions and magmatic oscillations can take on a variety of forms. Studies of the mechanisms of tremor excitation are still very much in their infancy, owing to the paucity of accurate and detailed seismic observations essential in documenting and analyzing such processes.

Source Processes of Very-Long-Period Signals

Unlike the LP signals, which are mainly interpreted as manifestations of acoustic resonance, VLP signals are viewed as the results of inertial forces associated with perturbations in the flow of magma and gases through conduits. An increasingly widespread use of broadband seismometers on volcanoes has led to an increasing number of observations, and VLP signals have by now been documented in volcanic areas all over the world, including Sakurajima (Kawakatsu et al. 1992, 1994), Unzen (Uhira et al. 1994), Aso (Kanesima et al. 1996, 2000; Legrand et al. 2000; Yamamoto et al. 1999), Satsuma-Iwojima (Ohminato and Ereditato 1997; Ohminato 2006), Iwate (Nishimura et al. 2000), Miyake-jima (Fujita et al. 2004; Kumagai et al. 2001), Usu

(Yamamoto et al. 2002), Bandai (Nishimura et al. 2003), Hachijo (Kumagai et al. 2003; Kumagai 2006), and Asama (Ohminato et al. 2006) in Japan; Stromboli (Auger et al. 2006; Chouet et al. 1999, 2003, 2008; Neuberg et al. 1994) in Italy; Merapi (Hidayat et al. 2000, 2002, 2003) in Indonesia; Kilauea (Almendros et al. 2002b, 2008; Dawson et al. 1998, 2004; Ohminato et al. 1998) in Hawaii; Long Valley (Hill et al. 2002; Hill and Prejean 2005) and Mount St. Helens (Waite et al. 2008) in the United States; Erebus (Aster et al. 2003; Rowe et al. 1998) in Antarctica; Popocatepetl (Arciniega-Ceballos et al. 1999, 2003; Chouet et al. 2005) in Mexico; and Cotopaxi and Tungurahua (Kumagai et al. 2007) in Ecuador.

The wavelengths of VLP signals are in the range of tens to hundreds of kilometers, which greatly facilitates their analysis. Inversions of VLP waveforms have imaged crack geometries at the source in the forms of dikes or sills (Chouet et al. 2003; Kumagai et al. 2003; Ohminato et al. 1998; Yamamoto et al. 1999), as well as more complicated geometrical configurations involving a composite of a dike intersecting a sill (Chouet et al. 2005), or composites of intersecting dikes (Almendros et al. 2008; Chouet et al. 2008), or two chambers connected to each other by a narrow channel (Nishimura et al. 2000). Based on the characteristic period of VLP signals recorded at Hachijo Island, Japan, Kumagai (2006) estimated a source with dimensions of a few km. The decaying harmonic oscillations with periods near 10 s and duration up to 300 s of the VLP signals seen at Hachijo are similar to the features of LP events with periods near 1 s (see Fig. 3) and appear consistent with the resonance of a dike containing a bubbly basalt. In that sense, these VLP signals probably represent a unique end member of the family of sources associated with resonant conduit excitation. Interestingly, the dominant periods and associated decay characteristics of the VLP signals at Hachijo were both found to vary with time, and these temporal variations were attributed by Kumagai (2006) to a change in source dimensions, assuming fixed acoustic properties of the fluid at the source. A spread of about 4 km was also estimated by

Chouet et al. (Nishimura et al. 2000) for two chambers at Iwate Volcano. The modeling of VLP waveforms at Stromboli by Nishimura et al. (Chouet et al. 2008) points to two point sources, each involving two intersecting dikes, whose relative positions and orientations represent individual conduit segments with dimensions of a few hundred meters. Systematic modeling of sustained VLP tremor accompanying episodes of deflation-inflation-deflation at Kilauea has suggested a complex magma pathway composed of a plexus of intersecting dikes extending over the depth range 0.6–1.6 km below the summit caldera (Almendros et al. 2008). A cracklike conduit extending over the depth range 0.3–2.5 km has also been inferred under Aso Volcano, Japan (Yamamoto et al. 1999).

Contrasting the above-discussed findings are results obtained by Ohminato et al. (2006) from waveform inversions of VLP signals produced by Vulcanian explosions at Asama Volcano, Japan. At Asama, the contribution from a vertical single force with magnitude 10^{10} – 10^{11} N was found to dominate the observed waveforms, and no obvious volumetric component was identified at the source. The depth of the VLP source is ~ 200 m beneath the summit crater, and the source-time history features two downward force components separated by an upward force component. The initial downward component was attributed by Ohminato et al. (2006) to the sudden removal of the lid capping the pressurized conduit (this force is analogous to the force F_{T_1} in Fig. 2), and the subsequent upward force was interpreted by these authors as a drag force induced by viscous magma moving up the conduit. Ohminato et al. (2006) also attributed the final downward force component to an explosive fragmentation of the magma, whose effect is to effectively cancel viscous drag so that the downward force due to jet recoil again dominates the VLP signal.

Available studies demonstrate that detailed investigations of the VLP oscillation characteristics of the source are critically important to our understanding of volcanic fluid dynamics. Below, I review two in-depth investigations of VLP signals representative of Vulcanian degassing bursts at Popocatepetl and gas-slug transport dynamics at Stromboli.

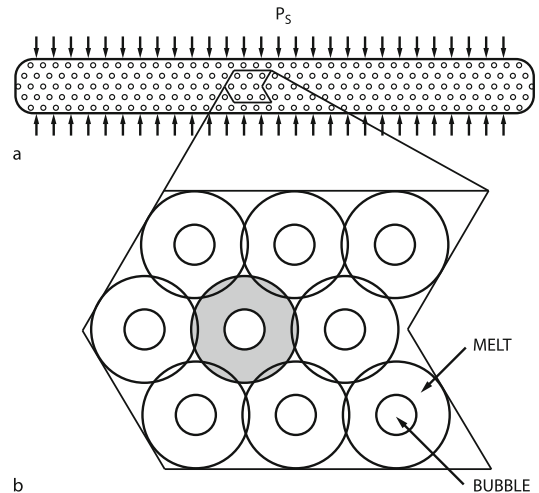
Coupled Diffusive-Elastic Pressurization at Popocatepetl Volcano

Vulcanian degassing bursts at Popocatepetl Volcano, Mexico, have been observed to be closely linked to the degassing of a sill-shaped volume of magma at a depth of 1.5 km below its summit crater (Chouet et al. 2005). The variations in volume of this sill were interpreted by Chouet et al. (2005) as reflecting cyclic pressure oscillations originating in the magma filling the sill. Based on these observations, these authors proposed a model in which static magma in the sill becomes supersaturated because of groundmass crystallization. In this conceptual view, volatile exsolution and diffusion of gas from the melt into bubbles increase the internal pressure of the bubbles, because bubble expansion is impeded by the viscous resistance of the surrounding liquid and by the confining effects due to the finite yield strength of the overlying column of magma and surrounding solid rock. Elastic inflation of the sill occurs as a result of bubble pressurization, and this inflation proceeds until the critical yield strength of the magma column is exceeded and magma starts flowing out of the sill. Magma fragmentation induced by viscous shear near the conduit wall then causes coalescence and collapse of bubbles intersected by fractures, allowing gas escape through a transient network of fractures. This, in turn, induces a pressure decrease in the sill, which results in the collapse and welding of the fracture network that shuts down the gas-escape pathway. Repeated cycles of shear-induced fracture and welding of magma provide a ratchet mechanism by which the separated gas phase in the magma can be recharged and evacuated.

The expansion dynamics of bubbles in supersaturated magma have been addressed in many previous studies under conditions of constant melt pressure (Lensky et al. 2002, 2004; Lyakhovsky et al. 1996; Proussevitch et al. 1993; Proussevitch and Sahagian 1998; Scriven 1959; Sparks 1978), or constant decompression rate (Proussevitch and Sahagian 1996; Toramaru 1989, 1995). A canonical model of bubble growth in magma that includes the effect of finite spacing of bubbles in the melt was first developed by

Proussevitch et al. (1993). This model considers a suspension of gas bubbles in an incompressible volatile-bearing liquid. The suspension is modeled as a three-dimensional lattice of closely-packed spherical cells, where each identical elementary cell is composed of a gas bubble surrounded by a shell of liquid. A step drop in pressure in the melt induces volatile exsolution and bubble expansion is driven by the diffusion of gas into the bubble. In this model, only the volatiles contained in the shell of liquid surrounding the bubble contribute to the bubble expansion so that bubble growth is limited by how closely packed the bubbles are in the liquid. Using the model of Proussevitch et al. (1993), Chouet et al. (2006) and Shimomura et al. (2006) obtained a dynamic solution that includes the effect of melt compressibility. In their model, pressure recovery is driven by bubble growth in a supersaturated magma that is stressed by the surrounding crust as the magma volume increases with bubble expansion. Both models consider the pressure recovery following an instantaneous pressure drop in the melt. The magma is embedded in an infinite, homogeneous, elastic solid and consists of a melt containing numerous small spherical gas bubbles of identical size. No new bubbles are created and no bubbles are lost during pressure recovery, and the gas in the bubbles is assumed to be a perfect gas. Gravity and other body forces are not considered.

The conduit geometry considered by Chouet et al. (2006) applies specifically to the observation made at Popocatepetl (Chouet et al. 2005) and consists of a penny-shaped crack containing a melt with an isotropic distribution of bubbles (Fig. 9a). The initial oversaturation of the melt resulting from a pressure drop ΔP_0 is distributed uniformly in the melt, and each bubble grows by diffusion of volatiles from a surrounding shell of melt with finite radius (Fig. 9b). As diffusion is a slow process, with time scale much longer than that of heat transfer into the bubble (Ichiara and Kameda 2004), bubble growth is assumed to be isothermal. The mathematical model describing an elementary cell therefore consists of three equations: (1) a diffusion equation governing the transfer of volatiles in the melt shell; (2) an equation



Volcanoes, Non-linear Processes in, Fig. 9 Model used by Chouet et al. (2006) to interpret the source mechanism of Vulcanian degassing bursts at Popocatepetl Volcano, Mexico. (a) Schematic illustration (cross section) of penny-shaped crack embedded in an elastic solid. The crack is under confining pressure P_s , and contains a melt with an isotropic distribution of gas bubbles of identical sizes. (b) Detailed view of the distribution of elementary cells composing the bubbly liquid (after Proussevitch et al. 1993). Each cell (as in gray-shaded example) consists of a gas bubble surrounded by a finite volume of melt. The elementary cells are organized in a three-dimensional lattice with slight overlap of the cells, where the volumes of intersecting melt and gaps are equal, so that the gas-volume fraction is $(R/S)^3$, where R is the bubble radius and S is the outer shell radius

expressing the radial motion of the bubble; and (3) a relation representing the ideal gas approximation. These equations are then subjected to three boundary conditions expressing the phase equilibrium at the bubble wall, mass flux at the bubble wall, and zero mass flux through the outer shell wall. As the bubble grows by mass transfer from the melt, melt is compressed and the surrounding rock matrix is deformed as a result. The pressure in the melt is balanced by the stress applied by the surrounding elastic medium, and in both Chouet et al. (2006) and Shimomura et al. (2006), this is formalized by assuming a quasistatic deformation of the magma conduit. Melt shrinkage from dehydration (Ochs and Lange 1999) is assumed to be negligible compared to melt compression due to bubble growth and is not considered.

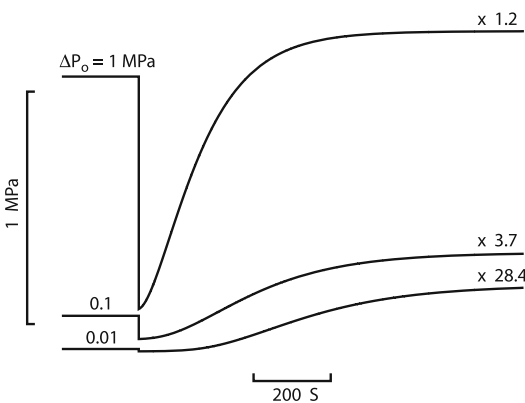
In their treatment of the initial condition in the bubble, Chouet et al. (2006) follow Proussevitch et al. (1993) and assume an instantaneous drop in gas pressure inside the bubble. This induces an instantaneous drop in volatile concentration at the bubble wall, thereby producing a step-like gradient of concentration that kick-starts the diffusion process. This condition is different from that assumed by Shimomura et al. (2006), who follow the model of Navon and Lyakhovsky (1998), in which the bubble initially retains the original gas pressure and immediately expands in response to the difference between internal and external pressures. The relaxation expansion of the bubble then induces a decrease in gas pressure inside the bubble, which leads to the establishment of a gradient in volatile concentration at the bubble wall that starts the diffusion. Although this leads to distinct features in the instantaneous bubble response, this has essentially no effect on the volumetric response of the bubble at longer time scales.

Figure 10 shows the pressure recovery calculated by Chouet et al. (2006) for input step drops $\Delta P_0 = 0.01, 0.1$, and 1 MPa applied to a bubbly rhyolitic melt encased in a penny-shaped crack with radius 100 m and thickness 5 m under ambient pressure $P_s = 40$ MPa appropriate for the depth of the source imaged for Popocatepetl Volcano (Chouet et al. 2005). The bubbles have a fixed initial radius 10^{-6} m, and the bubble number

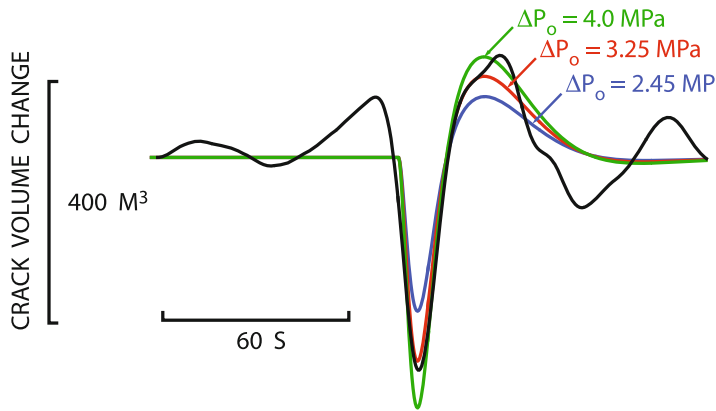
density is 10^{12} m^{-3} . Melt density, viscosity, and diffusivity are 2300 kg m^{-3} , 10^6 Pa s , and $10^{-11} \text{ m}^2 \text{ s}^{-1}$, respectively; surface tension is 0.2 Nm^{-1} , and the bulk modulus of melt and elastic rigidity of the rock are both 10^{10} Pa . The amplitude ratio of pressure recovery to initial pressure drop displays a strong nonlinear sensitivity to the magnitude of the input pressure drop, resulting in magnifications ranging from about 1 to nearly 30 over the range of input transients considered (Fig. 10). This overpressurization upon recovery was previously noted in a static solution obtained earlier by Nishimura (2004) and becomes largest for tiny bubbles and a stiff elastic medium (Chouet et al. 2006; Shimomura et al. 2006). Figure 10 also indicates that the speed of recovery is a function of the input transient, with the crack response becoming markedly slower for small pressure transients compared to larger ones.

The variation of time and amplitude scales seen in Fig. 10 suggests that a wide range of responses may be possible for cracks containing melts with a wider population of bubble sizes than the single-size bubble populations considered by Chouet et al. (2006). The net drop in volatile concentration associated with diffusion-driven bubble growth is quite small. Indeed, the results obtained by Chouet et al. (2006) point to concentration drops $\sim 0.003\text{--}0.035 \text{ wt}\%$ for initial bubble radii of $10^{-5}\text{--}10^{-7}$ m in penny-shaped cracks with aperture to radius ratios $0.01\text{--}0.25$. The net concentration drop is small because pressure recovery in the melt acts to raise the saturation level of volatiles in the melt. This has important implications for the overall history of degassing in a Vulcanian system, as it suggests that this process may be repeated many times without significantly depleting the gases in the melt body.

The source process imaged under Popocatepetl involves a sequence of inflation, deflation, and reinflation, reflecting a cycle of pressurization, depressurization, and repressurization within a time interval of $3\text{--}5$ min (Chouet et al. 2005). The volumetric component of the source processes in the sill associated with an eruption on 23 May 2000 is shown in Fig. 11. Other eruptions were found to produce similar waveform



Volcanoes, Non-linear Processes in,
Fig. 10 Dependence of pressure recovery on the size of the input transient, ΔP_0 , in the model of Chouet et al. (2006). The magnification factor for each response is indicated at the upper right of each pressure trace



Volcanoes, Non-linear Processes in, Fig. 11 Source-time function of volume change obtained by Chouet et al. (2005) for a sill under Popocatepetl Volcano during a degassing burst on 23 May 2000 (black line), and corresponding volume changes in a sill filled with a bubbly

melt in response to a step drop ΔP_0 in pressure (colored lines). The data from Popocatepetl and solutions from the crack model have both been band-pass filtered in the 15–70 s band

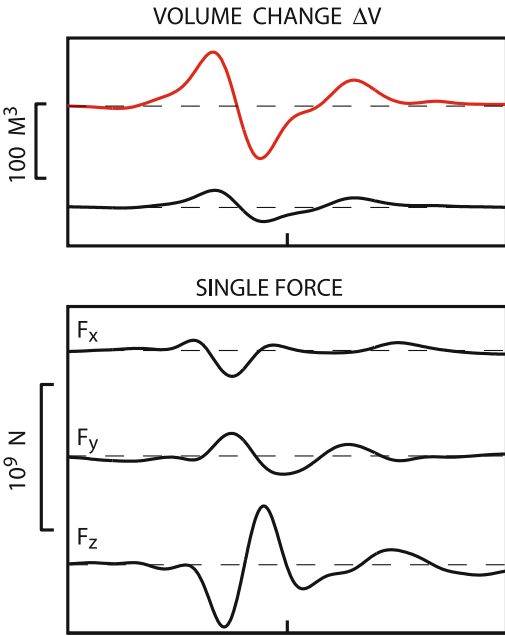
characteristics, hence the event in Fig. 11 may be viewed as an appropriate representation of overall source dynamics associated with these Vulcanian eruptions. For comparison, Fig. 11 also shows the volume change of a magma-filled sill calculated with the model of Chouet et al. (2006). The sill response has been band-pass filtered in the same band as the observed signal, and is illustrated for three distinct pressure drops. The model parameters in Fig. 11 are identical to those used in Fig. 10. A realistic fit of the peak-to-trough amplitude of the volume change imaged at Popocatepetl is obtained for an input pressure drop $\Delta P_0 = 3.25$ MPa.

Unfortunately, the longer-period components and other specific features of the sill response in the model are lost in the band-limited version of the signal compatible with observations made at Popocatepetl, so that the responses obtained in this band for different model parameters are all very similar (Chouet et al. 2006). Although it is not possible to infer specific values of melt viscosity, volatile diffusivity, initial bubble radius, bubble number density, or sill aperture to radius ratio based on available seismic data, these results strongly support the idea that diffusion pumping of bubbles in the magma under Popocatepetl provides a viable mechanism for pressure recovery following a pressure drop induced by a degassing event.

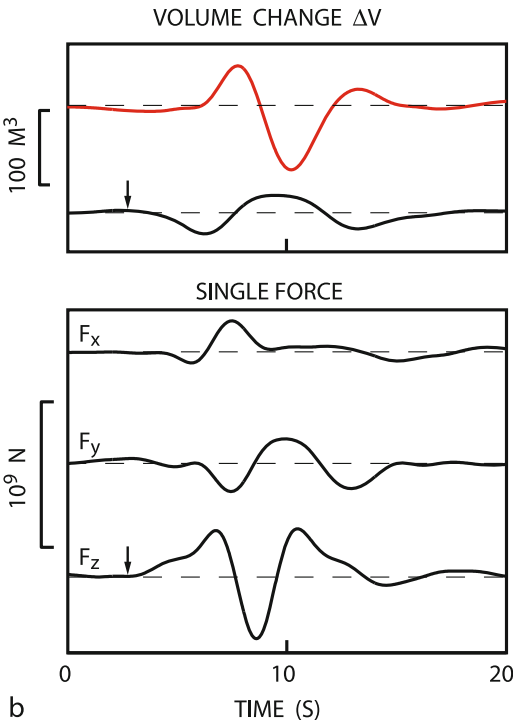
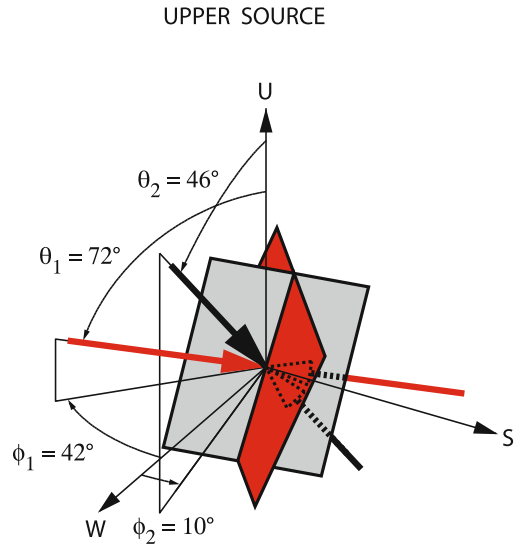
Slug Disruption at Stromboli

Eruptive behavior at Stromboli is characterized by mild, intermittent explosive activity, during which well-collimated jets of gases laden with molten lava fragments burst in short eruptions typically lasting 5–15 s. Modeling of VLP seismic data recorded during explosive activity in 1997 has imaged two distinct dike structures representative of explosive eruptions from two different vents located near the northern and southern perimeters of the summit crater (Chouet et al. 2003, 2008).

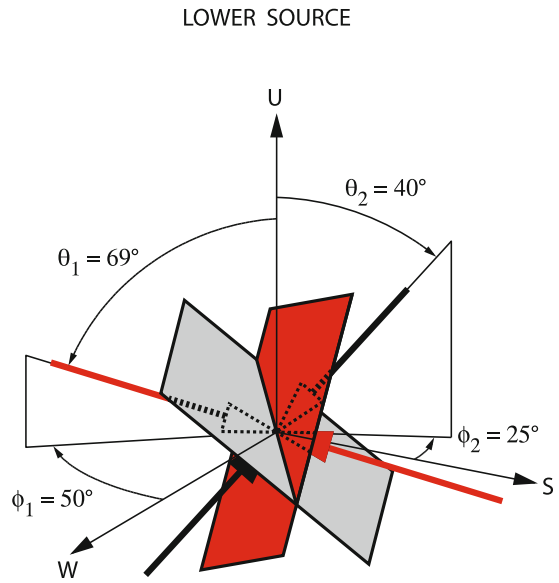
Figures 12 and 13 show two representations of the conduit geometry underlying the northern vent area. Figure 12 shows the seismic source mechanisms obtained from inversion of VLP waveforms; the mechanisms include two point sources, each of which marks a flow disruption site along the upper conduit. The upper source (Fig. 12a) is located 300 m below the crater floor and represents a bifurcation in the conduit; this is the main flow disruption site for gas slugs ascending toward the northern vent. The lower source (Fig. 12b), located roughly 500 m below the upper source, involves a sharp corner in the conduit and represents a secondary flow disruption site. Both upper and lower sources feature a dominant crack sustaining the largest volume change (colored red), and a subdominant crack undergoing a smaller volume change (shaded gray). Both cracks in Fig. 12a display a similar sequence of



a

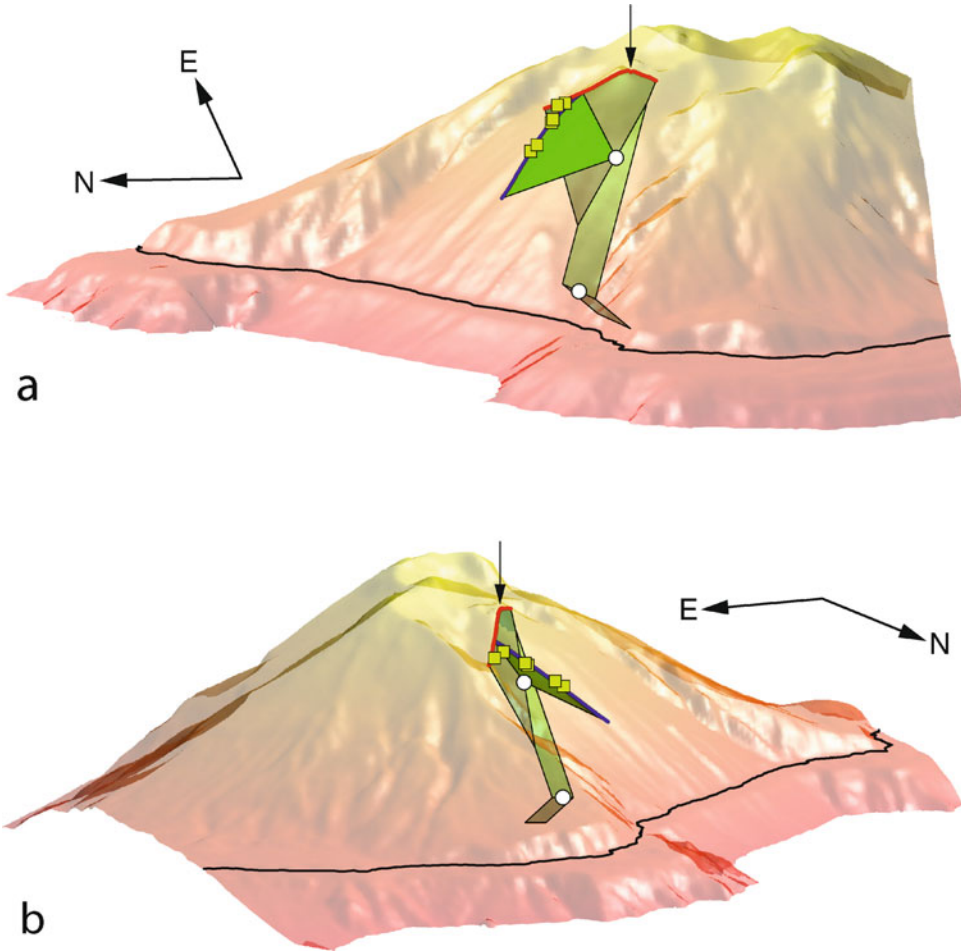


b



Volcanoes, Non-linear Processes in, Fig. 12 Seismic source mechanisms imaged for the two flow disruption sites in the shallow conduit structure underlying the northern vent area at Stromboli. The two point-source

mechanisms were obtained by Chouet et al. (2008) from inversions of VLP waveforms recorded during explosions at Stromboli, and are positioned at different depths in the volcanic edifice (see text for details). Each source consists



Volcanoes, Non-linear Processes in, Fig. 13 Geometry of the upper 1 km of conduit underlying the northern vent area of Stromboli. A semi-transparent view of the north-west quadrant of the volcanic edifice provides the reference for the location and geometry of the conduit, which is derived from the seismic source mechanisms in Fig. 12. Thin black line indicates sea level. The summit of the volcano is 924 m above sea level (no vertical exaggeration). The two flow disruption sites that are sources of VLP elastic radiation are indicated by small circles. The

irregular red and blue lines, respectively, represent the surface traces of the dominant and subdominant dike segments constituting the shallowest portions of the conduit system. The eruptive vent is marked by an arrow, and vents temporarily active during the flank eruption in 2002–2003 are marked by green squares. The lateral extents of individual dike segments are unknown and are shown for illustrative purpose only. (a) East-looking view. (b) South-looking view

Volcanoes, Non-linear Processes in, Fig. 12 (continued) of two intersecting cracks and a single force with components F_x (east), F_y (north), and F_z (up). Volume changes are color-coded with the color of the cracks they represent in each source. Crack orientations are provided by the azimuth ϕ and polar angle θ of the dominant dipole normal to the crack plane, with arrow directions marking

crack deflation (see Fig. 1c). The reference coordinates are W (west), S (south), and U (up). (a) Upper source. (b) Lower source. Small arrows in left panels mark the onset of deflation of the lower dike (gray dike in right panel) and synchronous start of the upward force. (Reproduced from Chouet et al. 2008)

inflation-deflation-inflation. The dominant crack in the lower source displays a volumetric response similar to that seen in the dominant crack at the upper source, but delayed by about 1 s with respect to the upper source. This implies a propagation speed of roughly 500 m/s between the two sources, consistent with the slow speed expected for the crack wave.

Figure 13 shows a picture of the upper conduit geometry consistent with the seismic source mechanisms imaged in Fig. 12. The closely matching dips of the two dominant cracks in Fig. 12 point to a conduit that extends essentially straight from 80 m below sea level to the crater floor, 760 m above sea level. At a depth of 80 m below sea level the conduit features a sharp corner leading into a dike segment dipping 40° to the southeast. The upper dominant dike segment, and deep segment below the abrupt corner both strike northeast-southwest along a direction parallel to the elongation of the volcanic edifice and a prominent zone of structural weakness, as expressed by lineaments, dikes, and brittle structures. The surface trace of the main dike segment trends through the northern vent area, while that of the upper subsidiary segment extends northwest-southeast in rough alignment with several vents active in the northwest quadrant of Stromboli in 2002–2003 (Acocella et al. 2006); the subsidiary dike trace intersects the main dike trace ~ 170 m north of the northern vent area.

A striking aspect of the mechanisms imaged for the two sources in Fig. 12 is the presence of dominantly vertical single-force components with common-looking time histories, except for a polarity reversal in one source compared to the other. The upper source (Fig. 12a) displays an initially downward force followed by an upward force, while an upward force followed by a downward force is manifest in the lower source (Fig. 12b). These force components compensate each other so that the total momentum in the overall source volume is conserved. In contrast to the delay of ~ 1 s in the volumetric components noted above, no significant delay is noted in the onsets of the vertical forces at the two sources, suggesting that transmission of the force between

the two sources occurs via the faster speed (3.5 km/s (Chouet et al. 2003)) of the compressional wave in the rock matrix.

Laboratory simulations carried out by James et al. (2006) provide insights into the origin of the initial pressurization and downward force observed at the upper source in Fig. 12a. These experiments investigate the ascent of a slug of gas in a vertical liquid-filled tube featuring a flare that abruptly doubles the cross sectional area. The tube is instrumented with pressure transducers mounted flush with the inner tube wall, and one accelerometer mounted on the exterior of the tube, and the whole assembly is free to move in the vertical direction. Detailed measurements of the flow transients obtained by James et al. (2006) show that the transit of a gas slug through the tube flare involves complex changes in flow pattern. A characteristic pinching of the slug tail is observed to occur synchronously with strong pressure and acceleration transients at the time the slug clears the flare, a picture consistent with the downward and inward motion of a liquid piston formed by the thickening film of liquid falling past the slug expanding in the wider tube. The sudden deceleration of the liquid annulus as it impinges the narrower inlet to the lower tube segment generates a pressure pulse in the liquid below the flare and also induces a downward force on the apparatus.

These observations are consistent with the pressurization phase and initial downward force imaged for the upper source at Stromboli and a similar funneling mechanism was inferred by Chouet et al. (2008) to be operative there. The repeatability of recorded pressure data and dependence of the magnitude of the pressure transient on slug size seen in the experiments of James et al. (2006) are also in harmony with the observed spatio-temporal properties of VLP signals at Stromboli (Chouet et al. 2003, 2008).

At the lower source, the start of the vertical force signal is synchronous with the onset of deflation of the lower dike segment (see arrows in the left panels in Fig. 12b). As the amplitude of the upward force increases, the lower dike segment continuously deflates. During the same interval, the dominant dike (red-colored

volumetric trace in Fig. 12b) remains in a slightly deflated state. The lower dike reaches maximum deflation at the time the upper dike segment goes through a transition from weak contraction to expansion, and the upward force reaches its peak amplitude ~ 0.5 s later. This picture is consistent with a compression of the lower dike synchronous with a downward acceleration of the liquid mass, both of which are suggestive of increasing external pressure on the conduit wall resulting from the downward vertical force acting at the upper source. Compression of the lower dike segment proceeds unimpeded until this process is overprinted by the arrival of the much slower volumetric expansion signal from the upper source.

Although not illustrated here, a similar slug disruption mechanism was imaged in the dike system underlying the southern vent (Chouet et al. 2008). In both conduit systems, the early response of the lower source relative to the volumetric disturbance arriving from the upper source may be interpreted as the passive response of the liquid to the movement of the conduit wall induced by elastic radiation from the force acting at the upper source. The scenario emerging from these dynamics is that of an upper source representing an active fluid phase and passive solid phase, and a lower source representing an active solid phase and passive fluid phase. The overall seismic source process associated with eruptions at Stromboli may then be summarized as follows. A slug of gas formed in the deeper reaches of conduit (Burton et al. 2007) rises through the lower conduit corner (the lower seismic source). At this point, the slug is most likely a few meters long and traverses this corner aseismically; past this corner, the slug expands on its way to the upper conduit bifurcation (the upper seismic source). The transit time from the lower to the upper seismic source is probably in the range of 5–15 min (James et al. 2006), hence related changes in magmatic head are well beyond the capability of broadband seismometers to detect and are not apparent within the VLP band imaged in Fig. 12. As it traverses the upper seismic source, the slug length has expanded to tens of meters and the slug is by then seismically noisy. Gravitational slumping of the liquid occurs as the slug expands through a flare in the conduit at this

location. The slumping liquid rapidly decelerates in the narrowing dike neck, increasing the liquid pressure and inducing a volume expansion in the main conduit and its subsidiary branch. The rapid deceleration and associated pressurization of the liquid couples to the conduit wall via the flare shoulders and induces a downward vertical force on the Earth. The volumetric signal propagates along the conduit at the slow speed of the crack wave, while the force signal itself propagates in the solid at the much higher speed of the compressional wave in the rock matrix, arriving at the lower conduit corner well before the crack wave. At this corner, the downward displacement of the rock induced by the force acting at the upper source impinges the bottom dike and squeezes this segment of conduit; this segment essentially acts like a spring that absorbs the downward motion of the rock. This scenario is consistent with both the small volume change of the lower dike segment, as well as its early response. The subsequent conduit response then reflects the combined effects of volumetric and mass oscillations of this liquid/gas/solid system, which are damped and eventually terminated by changing flow conditions.

The analyses carried out by Chouet et al. (2008) illuminate the subsurface processes driving eruptions at Stromboli and clearly point to the key role played by the conduit geometry in controlling fluid motion and resultant processes. Each irregularity or discontinuity in the conduit provides a site where pressure and momentum changes resulting from flow processes associated with the transit of a gas slug through the discontinuity are coupled to the Earth, or where the elastic response of the conduit can couple back into pressure and momentum changes in the fluid. The resulting processes are naturally oscillatory and involve diverse dynamics that reflect the complex physicochemical behavior of the volcanic system from the surface downward.

Future Directions

Seismology alone cannot directly see into the conduit and resolve details of the actual fluid dynamics at the origin of the seismic source

mechanisms revealed by analyses of LP or VLP signals. To develop a better understanding of fluid behavior responsible for these signals, laboratory experiments are required to explore the links between known flow processes and the resulting pressure and momentum changes. The pressure and momentum changes generated under laboratory conditions may then be compared with the pressure and momentum changes estimated from the time-varying moment-tensor and single-force components imaged from seismic data, yielding clues about the physical flow processes linked to the seismic source mechanism. Recent laboratory studies (James et al. 2004, 2006; Lane et al. 2001) have elucidated self-excitation mechanisms inherent to the fluid nonlinearity that are providing new insights into the mechanisms imaged from seismic data. In particular, the results obtained by James et al. (2006) demonstrate that direct links between the moment-tensor and single-force seismic-source mechanism and fluid-flow processes are possible and could potentially provide a wealth of information not available from seismic data alone. Together with these laboratory advances, numerical studies of multiphase flows are required to shed light on both the micro- and macrophysics of such flows (e.g., Badalassi et al. 2003), along with models exploring the coupled dynamics of fluid and solid (Nishimura and Chouet 2003). The key to a better understanding of volcanic processes lies in a sustained effort aimed at cross-fertilization between increasingly realistic numerical and experimental models of the fluid dynamics and elastodynamics, spatially and temporally dense field measurements of diverse geophysical signals at all frequencies, and chemical and physical evidence recorded in the eruptive products. In the next section, I present a brief summary of a modeling approach to two-phase fluids that holds great promise for the quantification of volcanic and hydrothermal source processes.

Phase-Field Method

Liquid-gas mixtures are an ubiquitous feature of magmatic and hydrothermal systems. A major challenge in addressing these types of fluids is the description of the moving and deforming interface between the two components of fluid. Traditional fluid dynamics treats these as sharp interfaces on which matching boundary conditions must be

imposed, which generally leads to intractable problems. Another difficulty is posed by the complex rheology of each component, whose internal microstructure is coupled with the flow field. The *phase-field method*, also known as the *diffuse-interface model*, circumvents these difficulties by considering the interface between components as a thin diffuse layer within which the two components are mixed and store a mixing energy. This idea follows from the original work of van der Waals (1979) and expresses the properties of the interface by molecular forces and a mixing energy. In the limit of an interface width approaching zero, the diffuse-interface model reduces to the classical sharp-interface model and results in the proper expression for interfacial tension. Lowengrub and Truski-novsky (1998) provide a detailed formulation of the conservative dynamics of the diffuse-interface model based on the classical procedure of Lagrangian mechanics. A review of diffuse-interface methods and related applications can also be found in Anderson et al. (1998). Noteworthy are the more recent applications by Kendon et al. (2001), Xu et al. (2003), Xu et al. (2004), and Yue et al. (2004). The following discussion is restricted to a brief description of the basic principles of the method based on the work of Kendon et al. (2001), who consider the evolution of a symmetric binary fluid mixture after a deep quench.

Equilibrium State A two-phase mixture is described by two scalar fields n_A and n_B expressing the local molar densities of the fluid components A and B . The corresponding local density, ρ , is given by:

$$\rho = m_A n_A + m_B n_B, \quad (2)$$

where m_A and m_B are the molecular weights of phases A and B , respectively. The local composition of the fluid is quantified by a dimensionless parameter, ϕ ,

$$\phi = \frac{n_A - n_B}{n_A + n_B}, \quad (3)$$

called the order parameter; the values of ϕ span the range $[-1, +1]$ with the value $+1$ denoting the pure phase A and the value -1 denoting the pure phase B .

The chemical equilibrium between the two phases is described through a minimization of the free energy $\mathcal{F}$, which represents the integral over the body of the local free-energy density (with units of Jm^{-3} or, equivalently, Nm^{-2}), a function of the fluid composition and its gradient (Cahn and Hilliard 1958):

$$\mathcal{F} = \int \left\{ f_0(\phi, P) + \frac{1}{2}k(\nabla\phi)^2 \right\} \text{d}\mathbf{r}. \quad (4)$$

The term $f_0(\phi, P)$ in the above equation represents the bulk free-energy density, which is assumed to take the form of a double-well with respect to ϕ . The detailed functional form of the double-well potential is not important (Bray 1994; Kendon et al. 2001; Swift et al. 1995, 1996; Xu et al. 2003, 2004; Yue et al. 2004) and may be set as (Kendon et al. 2001; Xu et al. 2003):

$$f_0(\phi, P) = \frac{a}{2}\phi^2 + \frac{b}{4}\phi^4 + P \ln \frac{P}{P_0}, \quad (5)$$

where the polynomial terms represent the bulk properties of the fluid. The parameter $a < 0$ and parameter b is always positive (Xu et al. 2003). The term in P yields a positive background pressure and does not affect the phase behavior (see discussion of pressure below). The form of this latter term is selected here for convenience, with P_0 representing an arbitrary reference pressure. As defined above, the function f_0 features two minima for $\phi = \pm\sqrt{-a/b}$ that correspond to the coexisting pure bulk phases. Assuming $b = -a$, one obtains the equilibrium values $\phi = \pm 1$ (Xu et al. 2003).

The squared gradient term in Eq. (4), with k a positive constant, represents weakly non-local interactions between the components. The effect of these interactions favors mixing of the components in contrast to the tendency of the bulk free energy f_0 , which is toward total separation of the phases into domains of pure components (Yue et al. 2004).

According to the model represented by Eq. (5), minimization of the free energy is achieved through the creation of two bulk domains at compositions $\phi = \pm 1$ separated by interfaces across

which the composition varies smoothly from one phase to the other. The change in $\mathcal{F}$ induced by a small local change in composition is described by the chemical potential, μ ,

$$\mu = \frac{\delta\mathcal{F}}{\delta\phi}, \quad (6)$$

and equilibrium between the coexisting bulk phases is reached when μ is everywhere zero (Cahn and Hilliard 1958; Kendon et al. 2001). Using Eqs. (4) and (5), one obtains the following expression for μ :

$$\mu = a\phi + b\phi^3 - \kappa\nabla^2\phi. \quad (7)$$

The equilibrium interfacial profile is given by $\phi(x) = \tanh(2x/\xi)$, where x is the coordinate normal to the interface (Xu et al. 2003). With the choice $b = -a$, this yields an interfacial width, $\xi = 2\sqrt{-2k/a}$, and surface tension, $\sigma = (2/3)\sqrt{-2ak}$ (Xu et al. 2003). The interface thickness may then be defined as the width over which 90% of the variation of ϕ occurs; this yields an equilibrium interface thickness of 1.47222ξ .

As interfaces in the fluid can exert non-isotropic forces, pressure is a tensor. The thermodynamic pressure tensor P_{ij}^{th} is obtained from the free energy as (Evans 1979; Kendon et al. 2001; Swift et al. 1995)

$$P_{ij}^{\text{th}} = \left(P \frac{\delta\mathcal{F}}{\delta P} + \phi \frac{\delta\mathcal{F}}{\delta\phi} - \left[f_0(\phi, P) + \frac{1}{2}k(\nabla\phi)^2 \right] \right) \delta_{ij} + k \frac{\partial\phi}{\partial x_i} \frac{\partial\phi}{\partial x_j} = P + P_{ij}^{\text{chem}}, \quad (8)$$

where P is the bulk pressure, and P_{ij}^{chem} is the chemical pressure component given by

$$P_{ij}^{\text{chem}} = \left[\frac{1}{2}a\phi^2 + \frac{3}{4}b\phi^4 - k\phi\nabla^2\phi - \frac{1}{2}k(\nabla\phi)^2 \right] \delta_{ij} + k \frac{\partial\phi}{\partial x_i} \frac{\partial\phi}{\partial x_j}, \quad (9)$$

in which δ_{ij} is the Kronecker symbol ($\delta_{ij} = 0$ for $i \neq j$, and $\delta_{ij} = 1$ for $i = j$). Note that the expression between brackets contributes to the isotropic

fluid pressure and that only the last term is anisotropic.

Conservation Equations The temporal evolution of the fluid composition ϕ is described by the Cahn-Hilliard diffusion-advection equation (Cahn and Hilliard 1959)

$$\frac{\partial \phi}{\partial t} + \mathbf{v} \cdot \nabla \phi = \nabla \cdot (\gamma \nabla \mu), \quad (10)$$

where γ is the order-parameter mobility and $\mathbf{v}(\mathbf{r})$ is the fluid velocity.

The dynamics of the fluid are described by the Navier–Stokes equations, whose exact forms depend on the system considered. The equations describing a viscous, compressible, isothermal flow are the conservation of mass:

$$\frac{\partial \rho}{\partial t} + \nabla \cdot (\rho \mathbf{v}) = 0, \quad (11)$$

and conservation of momentum:

$$\rho \left[\frac{\partial \mathbf{v}}{\partial t} + (\mathbf{v} \cdot \nabla) \mathbf{v} \right] = -\nabla \cdot \mathbf{P}^{\text{th}} + \nabla \cdot \boldsymbol{\tau} + \rho \mathbf{g}, \quad (12)$$

where $\mathbf{g}$ is the acceleration of gravity vector, $\mathbf{P}^{\text{th}}$ is the thermodynamic pressure tensor with components given by Eqs. (8) and (9), and $\boldsymbol{\tau}$ is the viscous stress tensor, which for a Newtonian fluid is given by

$$\tau_{ij} = \lambda e_{kk} \delta_{ij} + 2\eta e_{ij}, \quad (13)$$

where η is the coefficient of viscosity (dynamic viscosity) and λ is the second coefficient of viscosity generally defined as $\lambda = - (2/3)\eta$ (Tannehill et al. 1997). Summation over repeated indices is implied. In this equation e_{ij} is the strain rate tensor given by:

$$e_{ij} = \frac{1}{2} \left(\frac{\partial v_i}{\partial x_j} + \frac{\partial v_j}{\partial x_i} \right). \quad (14)$$

The field variables are the order parameter ϕ , the three components of fluid velocity $\mathbf{v}$, density ρ ,

and bulk pressure P . Under isothermal conditions P may be specified through a van der Waals equation of state. The diffusion-advection Eq. (10), continuity Eq. (11), and the three momentum Eq. (12), along with an equation of state for P , therefore provide a complete description of the physicochemical evolution of the binary fluid mixture. All other quantities can be obtained from the relations (2), (3), (7), (8), (9), (13), and (14). For non-isothermal flow, a complete description is obtained from Eqs. (10, 11 and 12), along with two additional equations describing the conservation of energy and entropy production (Anderson et al. 1998).

The thermodynamic properties of the two-phase system enter the above equations through the chemical potential $\mu = \delta \mathcal{F} / \delta \phi$, and pressure tensor $\mathbf{P}^{\text{th}}$, which are both obtained from the free energy Eq. (4). As a result, interfaces between the different fluid phases appear naturally within the model and do not need to be put in as boundary conditions. The procedure is quite general and may conceivably be applied to any complex fluid with a properly defined free energy.

The coupled Cahn–Hilliard–Navier–Stokes system of equations representing binary fluid systems is analytically intractable, and efforts have mainly been directed toward the development of robust, stable numerical schemes. The description of the fluid in terms of a smooth variation of the composition $\phi(\mathbf{r})$ provides a coarse-grained representation of the fluid where the smallest length scale is larger than the average distance between molecules. Under dynamical conditions, care must be taken to ensure that this coarse-grained description adequately represents the underlying microscopic physics. For the nominal interfacial thickness representing 90% of the variation of ϕ this typically requires 7–10 grids to resolve (Yue et al. 2004). Numerical applications include the Lattice Boltzmann Method (LBM) (Balazs et al. 2005; Kendon et al. 2001; Orlandini et al. 1995; Pooley et al. 2005; Swift et al. 1995, 1996; Xu et al. 2003, 2004), and various finite-difference approaches based on an Implicit-Explicit (IMEX) discretization (Ascher et al. 1995), semi-implicit discretization (Badalassi et al. 2003; Yue et al. 2004), fully implicit discretization (Kim et al.

2004), or fully explicit central-differenced staggered-grid discretization (Jacqmin 1999).

An example of application of this method that may provide a useful metaphor for the process of Vulcanian degassing discussed in section “Coupled Diffusive-Elastic Pressurization at Popocatepetl Volcano” is the 3D spinodal decomposition of a density-matched binary fluid mixture in a channel under shear (Badalassi et al. 2003). A remarkable feature of this type of flow is the formation of string-like structures. Similar structures have been observed in immiscible viscoelastic systems subjected to complex flow fields (Migler 2001), and such process may also play a role in the formation of degassing channels in a rhyolitic bubbly magma driven by slow pressurization in response to crystallization and de-gassing. A 2D application of the method to the dynamics of a gas slug ascending in a liquid-filled conduit featuring a flare is described in the chapter by *D’Auria and Martini*. The numerical results obtained by these authors reproduce the basic behavior observed in the experimental simulations of James et al. (2004) and James et al. (2006) and hold great potential as a key link between the seismic source mechanisms imaged at Stromboli (see section “Slug Disruption at Stromboli”) and the underlying volcanic fluid dynamics.

Acknowledgment I am grateful to Phil Dawson for his assistance in drafting figures. I am indebted to Robert Tilling and David Hill for careful reviews and helpful suggestions.

Bibliography

- Acocella V, Neri M, Scarlato P (2006) Understanding shallow magma emplacement at volcanoes: orthogonal feeder dykes during the 2002–2003 Stromboli (Italy) eruption. *Geophys Res Lett* 33:L17310. <https://doi.org/10.1029/2006GL026862>
- Aki K, Richards PG (1980) Quantitative seismology. Freeman, New York, p 932
- Aki K, Fehler M, Das S (1977) Source mechanism of volcanic tremor: fluid-driven crack models and their application to the 1963 Kilauea eruption. *J Volcanol Geotherm Res* 2:259–287
- Almendros J, Chouet B, Dawson P (2001) Spatial extent of a hydrothermal system at Kilauea Volcano, Hawaii, determined from array analyses of shallow long-period seismicity. *J Geophys Res* 106:13581–13597
- Almendros J, Chouet B, Dawson P, Huber C (2002a) Mapping the sources of the seismic wave field at Kilauea Volcano, Hawaii, using data recorded on multiple seismic antennas. *Bull Seismol Soc Am* 92:2333–2351
- Almendros J, Chouet B, Dawson P, Bond T (2002b) Identifying elements of the plumbing system beneath Kilauea Volcano, Hawaii, from the source locations of very-long-period signals. *Geophys J Int* 148:303–312
- Almendros J, Chouet B, Dawson P (2008) Shallow magma transport pathway under Kilauea Volcano, Hawaii, imaged from waveform inversions of very-long-period seismic data. 2-Source modeling. *J Geophys Res* (submitted)
- Anderson DM, McFadden GB, Wheeler AA (1998) Diffuse-interface methods in fluid mechanics. *Annu Rev Fluid Mech* 30:139–165
- Arciniega-Ceballos A, Chouet BA, Dawson P (1999) Very-long-period signals associated with Vulcanian explosions at Popocatepetl Volcano, Mexico. *Geophys Res Lett* 26:3013–3016
- Arciniega-Ceballos A, Chouet B, Dawson P (2003) Long-period events and tremor at Popocatepetl Volcano (1994–2000) and their broadband characteristics. *Bull Volcanol* 65:124–135
- Arciniega-Ceballos A, Chouet B, Dawson P, Asch G (2008) Broadband seismic measurements of degassing activity associated with lava effusion at Popocatepetl Volcano, Mexico. *J Volcanol Geotherm Res* 170:12–23
- Ascher UM, Ruuth SJ, Wetton BTR (1995) Implicit-explicit methods for time-dependent partial differential equations. *SIAM J Numer Anal* 3:797–823
- Aster R, Mah S, Kyle P, McIntosh W, Dunbar N, Johnson J, Ruiz M, McNamara S (2003) Very long period oscillations of Mount Erebus Volcano. *J Geophys Res* 108(B11). <https://doi.org/10.1029/2002JB002101>
- Auger E, D’Auria L, Martini M, Chouet B, Dawson P (2006) Real-time monitoring and massive inversion of source parameters of very long period seismic signals: an application to Stromboli Volcano, Italy. *Geophys Res Lett* 33:L04301. <https://doi.org/10.1029/2005GL024703>
- Badalassi VE, Cenicerros HD, Banerjee S (2003) Computation of multiphase systems with phase field models. *J Comput Phys* 190:371–397
- Balazs AC, Verberg R, Pooley CM, Kuksenok O (2005) Modeling the flow of complex fluids through heterogeneous channels. *Soft Matter* 1:44–54
- Bray AJ (1994) Theory of phase-ordering kinetics. *Adv Phys* 43:357–459
- Burton M, Allard P, Muré F, La Spina A (2007) Magmatic gas composition reveals the source depth of slug-driven Strom-bolian explosive activity. *Science* 317(5835): 227–230
- Cahn JW, Hilliard JE (1958) Free energy of a nonuniform system. I. Interfacial free energy. *J Chem Phys* 28: 258–267

- Cahn JW, Hilliard JE (1959) Free energy of a nonuniform system. III, Nucleation in a two-component incompressible fluid. *J Chem Phys* 31:688–689
- Cervelli PF, Miklius A (2003) The shallow magmatic system of Kilauea Volcano. *US Geol Surv Prof Pap* 1676: 149–163
- Chouet B (1986) Dynamics of a fluid-driven crack in three dimensions by the finite difference method. *J Geophys Res* 91:13967–13992
- Chouet B (1988) Resonance of a fluid-driven crack: radiation properties and implications for the source of long-period events and harmonic tremor. *J Geophys Res* 93: 4375–4400
- Chouet B (1992) A seismic source model for the source of long-period events and harmonic tremor. In: Gasparini P, Scarpa R, Aki K (eds) *IAVCEI proceedings in volcanology*, vol 3. Springer, New York, pp 133–156
- Chouet B (1996a) Long-period volcano seismicity: its source and use in eruption forecasting. *Nature* 380: 309–316
- Chouet B (1996b) New methods and future trends in seismological volcano monitoring. In: Scarpa R, Tilling RI (eds) *Monitoring and mitigation of volcano hazards*. Springer, New York, pp 23–97
- Chouet B (2003) Volcano seismology. *Pure Appl Geophys* 160(3–4):739–788
- Chouet B, Julian BR (1985) Dynamics of an expanding fluid-filled crack. *J Geophys Res* 90:11187–11198
- Chouet BA, Page RA, Stephens CD, Lahr JC, Power JA (1994) Precursory swarms of long-period events at Redoubt Volcano (1989–1990), Alaska: their origin and use as a forecasting tool. *J Volcanol Geotherm Res* 62:95–135
- Chouet B, Saccorotti G, Martini M, Dawson P, De Luca G, Milana G, Scarpa R (1997) Source and path effects in the wave fields of tremor and explosions at Stromboli Volcano, Italy. *J Geophys Res* 102:15129–15150
- Chouet B, Saccorotti G, Dawson P, Martini M, Scarpa R, De Luca G, Milana G, Cattaneo M (1999) Broadband measurements of the sources of explosions at Stromboli Volcano, Italy. *Geophys Res Lett* 26:1937–1940
- Chouet B, Dawson P, Ohminato T, Martini M, Saccorotti G, Giudicepietro F, De Luca G, Milana G, Scarpa R (2003) Source mechanisms of explosions at Stromboli Volcano, Italy, determined from moment-tensor inversions of very-long-period data. *J Geophys Res* 108(B1):2019. <https://doi.org/10.1029/2002JB001919>
- Chouet B, Dawson P, Arciniega-Ceballos A (2005) Source mechanism of Vulcanian degassing at Popocatepetl Volcano, Mexico, determined from waveform inversions of very long period signals. *J Geophys Res* 110: B07301. <https://doi.org/10.1029/2004JB003524>
- Chouet B, Dawson P, Nakano M (2006) Dynamics of bubble growth and pressure recovery in a bubbly rhyolitic melt embedded in an elastic solid. *J Geophys Res* 111:B07310. <https://doi.org/10.1029/2005JB004174>
- Chouet B, Dawson P, Martini M (2008) Shallow-conduit dynamics at Stromboli Volcano, Italy, imaged from waveform inversion. In: Lane SJ, Gilbert JS (eds) *Fluid motions in volcanic conduits a source of seismic and acoustic signals*. *Geol Soc Lond Spec Publ* 307: 57–84
- Collier L, Neuberg JW, Lensky N, Lyakhovsky V, Navon O (2006) Attenuation in gas-charged magma. *J Volcanol Geotherm Res* 153:21–36
- Commander KW, Prosperetti A (1989) Linear pressure waves in bubbly liquids: comparison between theory and experiments. *J Acoust Soc Am* 85:732–746
- Dawson PB, Dietel C, Chouet BA, Honma K, Ohminato T, Okubo P (1998) A digitally telemetered broadband seismic network at Kilauea Volcano, Hawaii. *US Geol Surv Open-File Rep* 98–108:1–121
- Dawson P, Whilldin D, Chouet B (2004) Application of near real-time radial semblance to locate the shallow magmatic conduit at Kilauea Volcano, Hawaii. *Geophys Res Lett* 31:L21606. <https://doi.org/10.1029/2004GL021163>
- Dzurisin D, Vallance JW, Gerlach TM, Moran SC, Malone SD (2005) Mount St. Helens reawakens. *Eos Trans AGU* 86:25, 29. <https://doi.org/10.1029/2005EO030001>
- Evans R (1979) The nature of the liquid-vapour interface and other topics in the statistical mechanics of non-uniform, classical fluids. *Adv Phys* 28:148–200
- Ferrazini V, Aki K (1987) Slow waves trapped in a fluid-filled infinite crack: implications for volcanic tremor. *J Geophys Res* 92:9215–9233
- Fujita E, Ida Y, Oikawa J (1995) Eigen oscillation of a fluid sphere and source mechanism of harmonic volcanic tremor. *J Volcanol Geotherm Res* 69:365–378
- Fujita E, Ukawa M, Yamamoto E (2004) Subsurface cyclic magma sill expansions in the 2000 Miyakejima Volcano eruption: possibility of two-phase flow oscillation. *J Geophys Res* 109:B04205. <https://doi.org/10.1029/2003JB002556>
- Gil Cruz F, Chouet BA (1997) Long-period events, the most characteristic seismicity accompanying the emplacement and extrusion of a lava dome in Galeras Volcano, Colombia, in 1991. *J Volcanol Geotherm Res* 77:121–158
- Hamada N, Jingu H, Ikumoto K (1976) On the volcanic earthquake with slowly decaying coda wave (in Japanese with English abstract). *Bull Volcanol Soc Jpn* 21:167–183
- Hasegawa A, Zhao D, Hori S, Yamamoto A, Horiuchi S (1991) Deep structure of the northeastern Japan arc and its relationship to seismic and volcanic activity. *Nature* 352:683–689
- Heliker C, Mattox T (2003) The first two decades of the Pu'u O'o-Kupaianaha eruption: chronology and selected bibliography. *US Geol Surv Prof Pap* 1676:1–27
- Hess K, Dingwell DB (1996) Viscosities of hydrous leucogranitic melts: a non-Arrhenian model. *Am Mineral* 81: 1297–1300
- Hidayat D, Voight B, Langston C, Ratdomopurbo A, Ebeling C (2000) Broadband seismic experiment at Merapi Volcano, Java, Indonesia: very-long-period

- pulses embedded in multiphase earthquakes. *J Volcanol Geotherm Res* 100:215–231
- Hidayat D, Chouet B, Voight B, Dawson P, Ratdomopurbo A (2002) Source mechanism of very-long-period signals accompanying dome growth activity at Merapi Volcano, Indonesia. *Geophys Res Lett* 29(23):2118. <https://doi.org/10.1029/2002GL015013>
- Hidayat D, Chouet B, Voight B, Dawson P, Ratdomopurbo A (2003) Correction to Source mechanism of very-long-period signals accompanying dome growth activity at Merapi Volcano, Indonesia. *Geophys Res Lett* 30(10):2118. <https://doi.org/10.1029/2003GL017211>
- Hill DP, Prejean S (2005) Magmatic unrest beneath Mammoth Mountain, California. *J Volcanol Geotherm Res* 146:257–283
- Hill DP, Dawson P, Johnston MJS, Pitt AM, Biasi G, Smith K (2002) Very-long-period volcanic earthquakes beneath Mammoth Mountain, California. *Geophys Res Lett* 29(10):1370. <https://doi.org/10.1029/2002GL014833>
- Hirabayashi J (1999) Formation of volcanic fluid reservoir and volcanic activity. *J Balneol Soc Jpn* 49:99–105
- Ichihara M, Kameda M (2004) Propagation of acoustic waves in a visco-elastic two-phase system: influences of the liquid viscosity and the internal diffusion. *J Volcanol Geotherm Res* 137:73–91
- Iverson RM, Dzurisin D, Gardner CA, Gerlach TM, LaHusen RG, Lisowski M, Major JJ, Malone SD, Messerich JA, Moran SC, Pallister JS, Qamar AI, Schilling SP, Vallance JW (2006) Dynamics of seismogenic volcanic extrusion at Mount St. Helens in 2004–05. *Nature* 444:439–443. <https://doi.org/10.1038/nature05322>
- Jacqmin D (1999) Calculation of two-phase Navier–Stokes Flows using phase-field modeling. *J Comput Phys* 155: 96–127
- James MR, Lane SJ, Chouet B, Gilbert JS (2004) Pressure changes associated with the ascent and bursting of gas slugs in liquid-filled vertical and inclined conduits. *J Volcanol Geotherm Res* 129:61–82
- James MR, Lane SJ, Chouet BA (2006) Gas slug ascent through changes in conduit diameter: laboratory insights into a volcano-seismic source process in low-viscosity magmas. *J Geophys Res* 111:B05201. <https://doi.org/10.1029/2005JB003718>
- Jones RAL (2002) *Soft condensed matter*. Oxford University Press, Oxford, p 195
- Julian BR (1994) Volcanic tremor, Nonlinear excitation by fluid flow. *J Geophys Res* 99:11859–11877
- Kanamori H, Given JW (1982) Analysis of long-period waves excited by the May 18, 1980, eruption of Mount St. Helens – a terrestrial monopole? *J Geophys Res* 87: 5422–5432
- Kanamori H, Given JW, Lay T (1984) Analysis of seismic body waves excited by the Mount St. Helens eruption of May 18:1980. *J Geophys Res* 89: 1856–1866
- Kaneshima S, Kawakatsu H, Matsubayashi H, Sudo Y, Tsutsui T, Ohminato T, Ito H, Uhira K, Yamasato H, Oikawa J, Takeo M, Iidaka T (1996) Mechanism of phreatic eruptions at Aso Volcano inferred from near-field broadband seismic observations. *Science* 273: 642–645
- Kawakatsu H (1989) Centroid single force inversion of seismic waves generated by landslides. *J Geophys Res* 94:12363–12374
- Kawakatsu H, Ohminato T, Ito H, Kuwahara Y (1992) Broadband seismic observation at Sakurajima Volcano, Japan. *Geophys Res Lett* 19:1959–1962
- Kawakatsu H, Ohminato T, Ito H (1994) 10-s-period volcanic tremors observed over a wide area in southwestern Japan. *Geophys Res Lett* 21:1963–1966
- Kawakatsu H, Kaneshima S, Matsubayashi H, Ohminato T, Sudo Y, Tsutsui T, Uhira K, Yamasato H, Ito H, Legrand D (2000) Aso94: Aso seismic observation with broadband instruments. *J Volcanol Geotherm Res* 101:129–154
- Kedar S, Sturtevant B, Kanamori H (1996) The origin of harmonic tremor at Old Faithful Geyser. *Nature* 379: 708–711
- Kendon VM, Cates ME, Pagonabarraga I, Desplat J-C, Bladon P (2001) Inertial effects in three-dimensional spinodal decomposition of a symmetric binary fluid mixture: a lattice Boltzmann study. *J Fluid Mech* 440: 147–203
- Kieffer SW (1977) Sound speed in liquid-gas mixtures: water-air and water-steam. *J Geophys Res* 82: 2895–2904
- Kim J, Kang K, Lowengrub J (2004) Conservative multi-grid methods for Cahn–Hilliard fluids. *J Comput Phys* 193:511–543
- Klein FW, Koyanagi RY, Nakata JS, Tanigawa WR (1987) The seismicity of Kilauea’s magma system. In: Decker RW, Wright TL, Stauffer RW (eds) *Volcanism in Hawaii*. US Geol. Surv. Prof. Pap., 1350. US Government Printing Office, Washington, DC, pp 1019–1185
- Koyanagi RY, Chouet B, Aki K (1987) Origin of volcanic tremor in Hawaii. Part I: Compilation of seismic data from the Hawaiian Volcano Observatory, 1972 to 1985. In: Decker RW, Wright TL, Stauffer RW (eds) *Volcanism in Hawaii*, US Geol. Surv. Prof. Pap., 1350. US Government Printing Office, Washington, DC, pp 1221–1257
- Kumagai H (2006) Temporal evolution of a magmatic dike system inferred from the complex frequencies of very long period seismic signals. *J Geophys Res* 111: B06201. <https://doi.org/10.1029/2005JB003881>
- Kumagai H, Chouet BA (1999) The complex frequencies of long-period seismic events as probes of fluid composition beneath volcanoes. *Geophys Int* 138:F7–F12
- Kumagai H, Chouet BA (2000) Acoustic properties of a crack containing magmatic or hydrothermal fluids. *J Geophys Res* 105:25493–25512
- Kumagai H, Ohminato T, Nakano M, Ooi M, Kubo A, Inoue H, Oikawa J (2001) Very-long-period seismic signals and caldera formation at Miyake Island, Japan. *Science* 293:687–690
- Kumagai H, Chouet BA, Nakano M (2002a) Temporal evolution of a hydrothermal system in Kusatsu-Shirane

- Volcano, Japan, inferred from the complex frequencies of long-period events. *J Geophys Res* 107(B10):2236. <https://doi.org/10.1029/2001JB000653>
- Kumagai H, Chouet BA, Nakano M (2002b) Waveform inversion of oscillatory signatures in long-period events beneath volcanoes. *J Geophys Res* 107(B11): 2301. <https://doi.org/10.1029/2001JB001704>
- Kumagai H, Miyakawa K, Negishi H, Inoue H, Obara K, Suet-sugu D (2003) Magmatic dyke resonances inferred from very-long-period seismic signals. *Science* 299:2058–2061
- Kumagai H, Chouet BA, Dawson PB (2005) Source process of a long-period event at Kilauea Volcano, Hawaii. *Geophys J Int* 161:243–254
- Kumagai H, Yepes H, Vaca M, Caceres V, Nagai T, Yokoe K, Imai T, Miyakawa K, Yamashina T, Arrais S, Vasconez F, Pinajota E, Cisneros C, Ramos C, Paredes M, Gomezjurado L, Garcia-Aristizabal A, Molina I, Ramon P, Segovia M, Palacios P, Troncoso L, Alvarado A, Aguilar J, Pozo J, EnriquezW MP, Hall M, Inoue I, Nakano M, Inoue H (2007) Enhancing volcano-monitoring capabilities in Ecuador. *EOS Trans Am Geophys Union* 88: 245–246
- Lane SJ, Chouet BA, Phillips JC, Dawson P, Ryan GA, Hurst E (2001) Experimental observations of pressure oscillations and flow regimes in an analogue volcanic system. *J Geophys Res* 106:6461–6476
- Legrand D, Kaneshima S, Kawakatsu H (2000) Moment tensor analysis of near-field broadband waveforms observed at Aso Volcano, Japan. *J Volcanol Geotherm Res* 101:155–169
- Lensky NG, Lyakhovsky V, Navon O (2002) Expansion dynamics of volatile-saturated liquids and bulk viscosity of bubbly magmas. *J Fluid Mech* 460:39–56
- Lensky NG, Navon O, Lyakhovsky V (2004) Bubble growth during decompression of magma: experimental and theoretical investigation. *J Volcanol Geotherm Res* 129:7–22
- Lowengrub J, Truskinovsky L (1998) Quasi-incompressible Cahn-Hilliard fluids and topological transitions. *Proc Roy Soc Lond A* 454:2617–2654
- Lyakhovsky V, Hurwitz S, Navon O (1996) Bubble growth in rhyolitic melts: experimental and numerical investigation. *Bull Volcanol* 58:19–32
- Major JJ, Scott WE, Driedger C, Dzurisin D (2005) Mount St. Helens erupts again – activity from September 2004 through March 2005. US Geological Survey Fact Sheet FS2005-3036. US Government Printing Office, Washington, DC, p 4
- Meier GEA, Grabitz G, Jungowsky WM, Witczak KJ, Anderson JS (1978) Oscillations of the supersonic flow downstream of an abrupt increase in duct cross section. *Mitteilungen aus dem Max-Planck Institut für Strömungsforschung und der Aerodynamischen Versuchsanstalt* 65:1–172
- Migler KB (2001) String formation in sheared polymer blends: coalescence, breakup, and finite size effects. *Phys Rev Lett* 86(6):1023–1026
- Moran SC, Malone SD, Qamar AI, Thelen W, Wright AK, Caplan-Auerbach J (2007) 2004–2005 seismicity associated with the renewed dome-building eruption of Mount St. Helens. In: Sherrod DR, Scott WE, Stauffer PH (eds) *A Volcano rekindled: the first year of renewed eruptions at Mount St. Helens, 2004–2006*. US Geological Survey Prof. Pap. 2007-XXXX
- Morrissey MM, Chouet BA (1997) A numerical investigation of choked flow dynamics and its application to the triggering mechanism of long-period events at Redoubt Volcano, Alaska. *J Geophys Res* 102: 7965–7983
- Nakano M, Kumagai H, Kumazawa M, Yamaoka K, Chouet BA (1998) The excitation and characteristic frequency of the long-period volcanic event: an approach based on an inhomogeneous autoregressive model of a linear dynamic system. *J Geophys Res* 103:10031–10046
- Nakano M, Kumagai H, Chouet BA (2003) Source mechanism of long-period events at Kusatsu-Shirane Volcano, Japan, inferred from waveform inversion of the effective excitation functions. *J Volcanol Geotherm Res* 122:149–164
- Navon O, Lyakhovsky V (1998) Vesiculation processes in silicic magmas. *Geol Soc Lond Spec Publ* 145:27–50
- Neuberg J, Luckett R, Ripepe M, Braun T (1994) Highlights from a seismic broadband array on Stromboli Volcano. *Geophys Res Lett* 21:749–752
- Neuberg JW, Tuffen H, Collier L, Green D, Powell T, Dingwell D (2006) The trigger mechanism of low-frequency earthquakes on Montserrat. *J Volcanol Geotherm Res* 153:37–50
- Nishimura T (2004) Pressure recovery in magma due to bubble growth. *Geophys Res Lett* 31:L12613. <https://doi.org/10.1029/2004GL019810>
- Nishimura T, Chouet B (2003) A numerical simulation of magma motion, crustal deformation, and seismic radiation associated with volcanic eruptions. *Geophys J Int* 153:699–718
- Nishimura T, Nakamichi H, Tanaka S, Sato M, Kobayashi T, Ueki S, Hamaguchi H, Ohtake M, Sato H (2000) Source process of very long period seismic events associated with the 1998 activity of Iwate Volcano, northeastern Japan. *J Geophys Res* 105:19135–19147
- Nishimura T, Ueki S, Yamawaki T, Tanaka S, Hasino H, Sato M, Nakamichi H, Hamaguchi H (2003) Broadband seismic signals associated with the 2000 volcanic unrest of Mount Bandai, northeastern Japan. *J Volcanol Geotherm Res* 119:51–59
- Ochs FA, Lange RA (1999) The density of hydrous magmatic liquids. *Science* 283:1314–1317
- Ohba T, Hirabayashi J, Nogami K (2000) D/H and $^{18}\text{O}/^{16}\text{O}$ ratios of water in the crater lake of Kusatsu-Shirane Volcano, Japan. *J Volcanol Geotherm Res* 97:329–346
- Ohminato T (2006) Characteristics and source modeling of broadband seismic signals associated with the hydrothermal system at Satsuma-Iwojima volcano, Japan. *J Volcanol Geotherm Res* 158:467–490
- Ohminato T, Chouet BA (1997) A free-surface boundary condition for including 3D topography in the finite-difference method. *Bull Seismol Soc Am* 87:494–515
- Ohminato T, Ereditato D (1997) Broadband seismic observations at Satsuma-Iwojima Volcano, Japan. *Geophys Res Lett* 24:2845–2848

- Ohminato T, Chouet BA, Dawson PB, Kedar S (1998) Waveform inversion of very-long-period impulsive signals associated with magmatic injection beneath Kilauea volcano, Hawaii. *J Geophys Res* 103: 23839–23862
- Ohminato T, Takeo M, Kumagai H, Yamashina T, Oikawa J, Koyama E, Tsuji H, Urabe T (2006) Vulcanian eruptions with dominant single force components observed during the Asama 2004 volcanic activity in Japan. *Earth Planets Space* 58:583–593
- Orlandini E, Swift MR, Yeomans JM (1995) A Lattice Boltzmann Model of binary fluid mixtures. *Europhys Lett* 32:463
- Pooley CM, Kuksenok O, Balasz AC (2005) Convection-driven pattern formation in phase-separating binary fluids. *Phys Rev E* 71:030501(R)
- Power JA, Stihler SD, White RA, Moran SC (2004) Observations of deep long-period (DLP) seismic events beneath Aleutian Arc Volcanoes, 1989–2002. *J Volcanol Geotherm Res* 138:243–266
- Proussevitch AA, Sahagian DL (1996) Dynamics of coupled diffusive and decompressive bubble growth in magmatic systems. *J Geophys Res* 101: 17447–17455
- Proussevitch AA, Sahagian DL (1998) Dynamics and energetics of bubble growth in magmas: analytical formulation and numerical modeling. *J Geophys Res* 103: 18223–18251
- Proussevitch AA, Sahagian DL, Anderson AT (1993) Dynamics of diffusive bubble growth in magmas: isothermal case. *J Geophys Res* 98: 22283–22307
- Ripepe M, Poggi P, Braun T, Gordeev E (1996) Infrasonic waves and volcanic tremor at Stromboli. *Geophys Res Lett* 23:181–184
- Rowe CA, Aster RC, Kyle PR, Schlue JW, Dibble RR (1998) Broadband recording of Strombolian explosions and associated very-long-period seismic signals on Mount Erebus Volcano, Ross Island, Antarctica. *Geophys Res Lett* 25:2297–2300
- Saccorotti G, Chouet B, Dawson P (2001) Wavefield properties of a shallow long-period event and tremor at Kilauea Volcano, Hawaii. *J Volcanol Geotherm Res* 109:163–189
- Scriven LE (1959) On the dynamics of phase growth. *Chem Eng Sci* 10:1–13
- Shaw HR, Chouet B (1989) Singularity spectrum of intermittent seismic tremor at Kilauea Volcano, Hawaii. *Geophys Res Lett* 16:195–198
- Shaw HR, Chouet B (1991) Fractal hierarchies of magma transport in Hawaii and critical self-organization of tremor. *J Geophys Res* 96:10191–10207
- Shimomura Y, Nishimura T, Sato H (2006) Bubble growth processes in magma surrounded by an elastic medium. *J Volcanol Geotherm Res* 155:307–322
- Sparks R (1978) The dynamics of bubble formation and growth in magmas: a review and analysis. *J Volcanol Geotherm Res* 3:1–38
- Stephens CD, Chouet BA (2001) Evolution of the December 14, 1989 precursory long-period event swarm at Redoubt Volcano, Alaska. *J Volcanol Geotherm Res* 109:133–148
- Swift MR, Osborn WR, Yeomans JM (1995) Lattice Boltzmann simulation of nonideal fluids. *Phys Rev Lett* 75(5):0–833
- Swift MR, Orlandini E, Osborn WR, Yeomans JM (1996) Lattice Boltzmann simulations of liquid-gas and binary fluid systems. *Phys Rev E* 54(5): 5041–5052
- Takei Y, Kumazawa M (1994) Why have the single force and torque been excluded from seismic source models? *Geophys J Int* 118:20–30
- Tannehill JC, Anderson DA, Pletcher RH (1997) Computational fluid mechanics and heat transfer, 2nd edn. Taylor and Francis, Philadelphia, p 792
- Toramaru A (1989) Vesiculation process and bubble size distributions in ascending magmas with constant velocities. *J Geophys Res* 94:17523–17542
- Toramaru A (1995) Numerical study of nucleation and growth of bubbles in viscous magmas. *J Geophys Res* 100:1913–1931
- Tuffen H, Dingwell DB, Pinkerton H (2003) Repeated fracture and healing of silicic magma generates flow banding and earthquakes? *Geology* 31:1089–1092
- Uhira K, Yamasato H, Takeo M (1994) Source mechanism of seismic waves excited by pyroclastic flows observed at Unzen Volcano, Japan. *J Geophys Res* 99: 17757–17773
- van der Waals JD (1979) The thermodynamic theory of capillarity under the hypothesis of a continuous variation of density. *Verhandel Konink. Acad. Weten. Amsterdam, (Sect. 1), 1*, pp 1–56. Translation by Rowlingson JS. *J Stat Phys* 20:197–244
- Vergnolle S, Brandeis G, Mareschal JC (1996) Strombolian explosions, 2. Eruption dynamics determined from acoustic measurements. *J Geophys Res* 101:20449–20466
- Waite GP, Chouet BA, Dawson PB (2008) Eruption dynamics at Mount St. Helens imaged from broadband seismic waveforms: interaction of the shallow magmatic and hydrothermal systems. *J Geophys Res* 113: B02305. <https://doi.org/10.1029/2007JB005259>
- Wallis GB (1969) One-dimensional two-phase flow. McGraw-Hill, New York, p 408
- White RA (1996) Precursory deep long-period earthquakes at Mount Pinatubo: spatio-temporal link to a basalt trigger. In: Newhall CG, Punongbayan RS (eds) *Fire and mud: eruptions and lahars of Mount Pinatubo*, Philippines. University of Washington Press, Seattle, pp 307–326
- Wilson L, Head JW (1981) Ascent and eruption of basaltic magma on the earth and moon. *J Geophys Res* 86: 2971–3001
- Xu A, Gonnella G, Lamura A (2003) Phase-separating binary fluids under oscillatory shear. *Phys Rev E* 67: 056105
- Xu A, Gonnella G, Lamura A (2004) Phase separation of incompressible binary fluids with lattice Boltzmann methods. *Phys A* 331:10–22
- Yamamoto M, Kawakatsu H, Kaneshima S, Mori T, Tsutsui T, Sudo Y, Morita Y (1999) Detection of

- a crack-like conduit beneath the active crater at Aso Volcano, Japan. *Geophys Res Lett* 26: 3677–3680
- Yamamoto M, Kawakatsu H, Yomogida K, Koyama J (2002) Long-period (12 sec) volcanic tremor observed at Usu 2000 eruption: seismological detection of a deep magma plumbing system. *Geophys Res Lett* 29(9):1329. <https://doi.org/10.1029/2001GL013996>
- Yoon SW, Crum LA, Prosperetti A, Lu NQ (1991) An investigation of the collective oscillations of a bubble cloud. *J Acoust Soc Am* 89:700–706
- Yue P, Feng JJ, Liu C, Shen J (2004) A diffuse-interface method for simulating two-phase flows of complex fluids. *J Fluid Mech* 515:293–317
- Zhang Y, Xu Z, Liu Y (2003) Viscosity of hydrous rhyolitic melts inferred from kinetic experiments: a new viscosity model. *Am Mineral* 88:1741–1752

Volcano Deformation: Insights into Magmatic Systems

Daniel Dzurisin
1300 S.E. Cardinal Court, U.S. Geological
Survey, David A. Johnston Cascades Volcano
Observatory, Vancouver, WA, USA

Article Outline

Glossary
Volcano Geodesy: The Science of Deforming
Volcanoes
Introduction to Volcanoes and Volcano Science
Seismology, Geochemistry, Geology, and
Geodesy: Four Pillars of Modern Volcano
Science
Techniques for Measuring Volcano Deformation
Modeling and Analysis Techniques
What Can Be Learned: Case Studies
Future Directions
Bibliography

Glossary

Assimilation Process of incorporation and digestion of material from wall rock into magma. The resulting magma is sometimes referred to as hybrid or contaminated.

Basalt Volcanic rock or lava that is characteristically dark in color and low in viscosity, contains 45 to 54 wt.% silica (SiO_2), and is rich in iron and magnesium. Basalt lavas are typically much lower in viscosity than rhyolite lavas.

Bayesian Refers to methods in probability and statistics named after Thomas Bayes (c. 1701–1761), an English statistician and philosopher who formulated Bayes' theorem,

which provides a means to update a prior probability distribution to account for new evidence.

Benchmark A relatively permanent, natural, or artificial geodetic marker fixed to the Earth and bearing a marked point whose coordinates in some specified reference system have been determined precisely. Sometimes written “bench mark.”

Brine Water containing a high concentration of one or more dissolved salts is called a brine. The solubilities of chloride-bearing salts generally increase as temperature and pressure increase deeper into the Earth, so brines containing much more than 25 wt.% dissolved salts occur there. Such hypersaline brines are likely to form where water-bearing fluids are released at very high temperatures and high pressures from crystallizing magmas a few to several kilometers deep beneath volcanoes.

Caldera A large, roughly circular or oval-shaped volcanic depression with a diameter many times larger than that of included volcanic vents and may range from 2 to 50 km across. Calderas form by collapse, erosion, or explosion. The largest calderas form when the roof of a magma chamber collapses during eruption of a large volume of silica-rich magma that forms an extensive ash-flow tuff. Examples of collapse calderas and associated tuffs in the United States include Crater Lake OR (Wineglass Tuff), Long Valley Caldera CA (Bishop Tuff), and Yellowstone Caldera WY (Lava Creek Tuff).

Cryptodome A shallow intrusion of viscous magma that typically results in a bulge or welt at the surface. A classic example is the dacite cryptodome that formed beneath the north-flank bulge at Mount St. Helens prior to the climactic eruption on May 18, 1980.

Deflation In volcanology, a term used to describe detumescence or shrinking of the ground surface, a volcanic edifice, or a subsurface magma body. Deflation is a response to depressuriza-

tion, which can have several causes including magma withdrawal (related to intrusion or eruption), volatile escape, thermal contraction, phase changes during crystallization, and tectonic extension.

Deformation Any change in the shape or dimensions of a body resulting from stress. In geodesy, the term generally refers to changes in the configuration of the ground surface (e.g., uplift, subsidence, extension, compression, shear) as a result of internal processes (e.g., faulting, magma intrusion) or external forces (e.g., tidal bulges).

Digital elevation model (DEM) A representation of the topography of the Earth in digital format, that is, by coordinates and numerical descriptions of altitude. Commonly used units are (longitude, latitude, elevation) and (x, y, z) in meters or feet.

Dike A tabular intrusion of magma that cuts across the bedding or foliation of surrounding rock. Dikes are distinguished from sills, which are concordant (parallel) with bedding or foliation. Dikes typically dip steeply, whereas sills dip gently. However, the dip of an intrusive body is not diagnostic: A vertical intrusion along a vertical bedding or foliation plane is a sill, and a horizontal intrusive body that cuts across vertical bedding or foliation is a dike.

Dilatometer Synonym for volumetric strainmeter. Two extremely precise borehole designs (sensitivity in the range 10^{-12} – 10^{-11} in volumetric strain) that have been used successfully for volcano monitoring are the Sacks-Evertson volumetric strainmeter and the Gladwin tensor strainmeter.

Dislocation In geology, the relative movement between the two sides of a fault or crack.

Elasticity Property of materials that deform reversibly under stress. Earth's crust is elastic under stresses that do not exceed its strength (elastic limit), at which point brittle or ductile failure occurs. Propagation of seismic waves over great distances is a demonstration of Earth's elasticity.

Electronic distance meter (EDM) A surveying device used to determine distance between fixed points by measuring the round-trip travel

time for transmitted electromagnetic signals from the instrument to a reflective target.

Fractionation (fractional crystallization)

Crystallization process in magmas in which early-formed crystals are prevented from equilibrating with the liquid from which they grew, resulting in a series of residual liquids with more extreme compositions than would have resulted from equilibrium crystallization.

Fumarole A volcanic vent from which gases and vapors are emitted. Fumaroles may occur along a fissure or in apparently chaotic clusters or fields.

Geodesy The science concerned with the determination of the size and shape of the Earth and the precise location of points on its surface. The scale of interest can range from global (Earth tides) to local (volcano deformation).

Global Positioning System (GPS) Acronym for the US Global Positioning System, a worldwide positioning and navigation tool that makes use of a constellation of NAVSTAR satellites in Earth orbit. GLONASS (operational), Galileo (in deployment phase, to be operational in 2020), and BeiDou (in deployment phase, to be operational globally as COMPASS in 2020) are comparable systems operated by Russia, the European Union, and China, respectively. Currently, NAVSTAR GPS, GLONASS, Galileo, and BeiDou comprise the Global Navigation Satellite System (GNSS). In common usage, GPS also refers more generally to the precise positioning technique that makes use of the GNSS.

GNSS Acronym for Global Navigation Satellite System, which in early 2019 includes the United States' NAVSTAR Global Positioning System (GPS), Russia's GLONASS, the European Union's Galileo, and China's BeiDou (COMPASS). Galileo and BeiDou (COMPASS) are in deployment phase and scheduled to be operational in 2020.

Hillshade The hypothetical illumination of a surface represented by shadows on a map that simulate the effect of the sun's rays and create a three-dimensional effect on the map.

Inflation In volcanology, a term used to describe tumescence (swelling) of the ground surface, a

volcanic edifice, or a subsurface magma body. Inflation is a response to pressurization, which can have several causes including magma accumulation, exsolution of volatiles, geothermal processes, heating, and tectonic compression.

Interferogram (radar) An image produced by differencing two radar-phase images of the same target area. An interferogram can be used to produce a digital elevation model, to portray ground deformation, or to detect other forms of surface change (e.g., erosion, deposition, or a change in near-surface moisture content).

Leveling A surveying technique for determining relative height differences between fixed points on Earth's surface (benchmarks) by sighting from a surveying instrument adjusted precisely to a horizontal attitude to one or more graduated rods. By repeating surveys that adhere to the strictest standards, elevation changes among a network of benchmarks can be measured to a precision $0.5 \text{ mm } \sqrt{L}$, where L is the distance in kilometers between marks. This corresponds to a theoretical precision of $\pm 1.6 \text{ mm}$ (one standard deviation) for marks separated by 10 km; in practice, the accuracy achieved depends on various factors that can be mitigated but not eliminated.

Lidar Remote-sensing method that uses light from a pulsed laser to measure ranges to the Earth surface or other target. Airplanes and helicopters are the most commonly used platforms for acquiring lidar data over broad areas. Ground-based lidars are used to scan smaller features at higher resolution. Topographic lidar typically uses a near-infrared laser to map the land, while bathymetric lidar uses water-penetrating green light to measure seafloor and riverbed elevations.

Plastic deformation Permanent change in shape or volume of a material without rupture, characterized by a yield stress that must be exceeded before deformation begins. Rocks and soils exhibit plastic behavior, i.e., once their elastic limit (strength) is exceeded, they fail either by flow (ductile deformation) or fracture (brittle deformation).

Plinian Refers to a type of explosive volcanic eruption in which a steady, turbulent stream of fragmented magma and magmatic gas is released at high velocity from a vent. Tall eruption columns, with heights 10–25 km above sea level or more, and volumes of erupted products in the range 10^7 – 10^{12} m^3 , are characteristic. Recent examples include the Mount St. Helens eruption of 18 May 1980 and the Mount Pinatubo eruption of 12–15 June 1991.

Poroelasticity Property of a material whose solid matrix is elastic and whose fluid is viscous. Fluid-saturated rock and soil are poroelastic materials.

Recharge As used here the term refers to re-accumulation of magma in a subsurface reservoir following a withdrawal event, which can be sudden or gradual. Synonymous with “resupply.”

Rhyolite Volcanic rock or lava that characteristically is light in color, contains 69–80 wt.% silica (SiO_2), and is rich in potassium and sodium. Rhyolite lavas are typically much higher in viscosity than basalt lavas.

Rift zone In a volcanic setting such as Hawai'i or Iceland, a linear zone of fractures, vents, and collapse features associated with an underlying dike complex. Rift zones are an important aspect of the magma plumbing system for many basaltic volcanoes. Magma that rises from depth to within a few kilometers of the surface can move laterally along a rift zone for tens of kilometers before erupting or freezing as an intrusion.

Sill A tabular igneous intrusion that is concordant (parallel) with the planar structure of the surrounding rock. Sills are distinguished from dikes, which cut across bedding or foliation. Sills typically dip gently, whereas dikes dip steeply. However, the dip of an intrusive body is not diagnostic: A vertical intrusion along a vertical bedding or foliation plane is a sill, and a horizontal intrusive body that cuts across vertical bedding or foliation is a dike.

Strainmeter An instrument designed to measure changes in the shape or volume of a body (deformation) as a result of stress. Two

extremely precise borehole designs (sensitivity in the range 10^{-12} – 10^{-11} in volumetric strain) that have been used successfully for volcano monitoring are the Sacks-Evertson volumetric strainmeter and the Gladwin tensor strainmeter.

Thermoelasticity As used here, the tendency for materials to expand as temperature increases or contract as temperature decreases. Thermoelastic deformation of porous rock can occur when pore fluid pressure changes as a result of heating or cooling.

Tiltmeter An instrument that measures slight changes in the tilt of the ground surface, usually in relation to a liquid-level surface or to the rest position of a pendulum.

Triangulation A surveying method for determining the directions and distances to, and the coordinates of, a network of points by means of bearings from two fixed points a known distance apart. Any relative movement among the points can be measured by repeating the survey at a later date.

Trilateration A surveying method in which the lengths of the three sides of a series of triangles that share common points at their vertices are measured and the angles between sides are computed from the measured lengths. Thus the coordinates of each point in the network can be established relative to a known point, and any relative movement among the points can be measured by repeating the survey at a later date.

Viscoelasticity Property of materials that exhibit both viscous and elastic behavior when undergoing deformation. When subjected to a constant stress, viscoelastic materials experience a time-dependent increase in strain (creep) until failure occurs. Rocks at high temperature exhibit viscoelastic behavior.

Together with seismicity and volcanic gas flux, deformation of the ground surface can be a key indicator of subsurface conditions and processes at volcanoes – information that not only improves scientific understanding of magmatic systems but also is useful for assessing volcano hazards and mitigating their potential consequences. To take full advantage of such information requires detailed characterization of the deformation field in space and time. Currently, no single geodetic technique can provide both the high spatial and temporal resolution required. However, important advances have been made recently by combining information from real-time in situ sensors such as continuous GPS, strainmeters, and tiltmeters with repeated campaign-style GPS, microgravity, interferometric synthetic-aperture radar (InSAR), lidar, and photogrammetric observations. Continuous real-time data constrain the timing but not necessarily the spatial extent and pattern of deformation, whereas InSAR provides detailed spatial information but only at intervals of several days to weeks. Repeated microgravity surveys, when combined with independent measurements of surface height, are uniquely sensitive to changes in subsurface mass distribution and therefore can be used to distinguish among processes driven by magma, hydrous fluids, or gas. Simultaneous analysis of multiple geodetic datasets, especially when guided by information from global volcano databases and numerical simulations of volcanic processes and products, can provide a statistical basis for outcome prediction and serve as a guide for additional observations. In the foreseeable future, interactions among magmatic, tectonic, and hydrothermal systems will be monitored and modeled at regional scale in real time, enabling new insights into Earth's subsurface environment.

Introduction to Volcanoes and Volcano Science

Volcano Geodesy: The Science of Deforming Volcanoes

Volcano geodesy is the branch of geodetic science that deals with the changing shapes of volcanoes, whether large or small, deep-seated or surficial.

Volcanoes have been a part of human experience from time immemorial. Ancient Romans believed volcanic eruptions were the work of Vulcan, a mythical blacksmith and the god of fire. Plato and Aristotle speculated on the causes of

eruptions more than 2300 years ago, attributing them to a subterranean river of fire or friction from the wind, respectively. Pliny the Elder noted the occurrence of earthquakes prior to the 79 CE eruption of Mount Vesuvius, and his nephew, Pliny the Younger, described a towering column of ash from the eruption – a type now referred to as “Plinian” to commemorate their contributions. The oral traditions passed down from ancient Hawaiians refer to Pele, the volcano deity, as the source of eruptions; in many respects the accounts are consistent with modern understanding of the pre-European eruptive history of the Hawaiian archipelago (Swanson 2008).

The beginning of the modern era of volcanology can be traced to the establishment in 1841 of the Osservatorio Vesuviano (Vesuvius Observatory) on the west flank of Mount Vesuvius, Italy. Today the historic structure serves as a museum, while scientists monitor Vesuvius from a modern facility in the nearby city of Naples. In 1903, the French established an observatory on the island of Martinique in the West Indies to monitor activity following the disastrous eruption of Montagne Pelée in 1902, which killed 29,000 people in the town of St. Pierre. A few hours earlier, an eruption at nearby La Soufrière volcano on the island of St. Vincent had killed 1500 people. Thomas A. Jaggar, an instructor at Harvard University and part-time employee of the US Geological Survey (USGS), was a member of the US scientific expedition sent to investigate the twin disasters. In 1909 Jaggar, then a professor at the Massachusetts Institute of Technology, traveled to Japan to visit renowned seismologist Fusakichi Omori. The two young scientists, who became lifelong friends, undoubtedly discussed the need for systematic long-term monitoring of dangerous volcanoes. Omori established the Asama Volcano Observatory, the first of several in Japan, 2 years later in 1911. Jaggar left MIT the following year to establish the Hawaiian Volcano Observatory (HVO) near the summit of Kīlauea Volcano on the Island of Hawai‘i. The Whitney Seismological Laboratory, precursor to the modern HVO, was constructed in accordance with plans drawn up by Omori, who also sent two of the first seismic monitoring instruments to be installed there. In

the span of a decade, two neighboring volcanoes in the Lesser Antilles and two close friends from a world apart had altered the course of volcanology forever.

Since the pioneering efforts by Jaggar and Omori a century ago, dozens of volcano observatories have been established worldwide. In the United States, HVO has been joined by four additional volcano observatories operated by the USGS with partner universities and government agencies: the Alaska Volcano Observatory (AVO), California Volcano Observatory (CalVO), Cascades Volcano Observatory (CVO), and Yellowstone Volcano Observatory (YVO) (<http://volcanoes.usgs.gov/>). In addition, the Volcano Disaster Assistance Program (VDAP), a joint undertaking of the USGS and the US Agency for International Development, responds to requests for assistance during volcano emergencies worldwide. According to the World Organization of Volcano Observatories (WOVO, <http://www.wovo.org/observatories/>), Japan now operates 16 volcano observatories and related research centers; other countries and regions with robust volcano research programs include Europe, Iceland, Indonesia, Latin America, the Lesser Antilles, New Zealand, Papua New Guinea, Republic of the Philippines, and Russia, among others.

In January 2019, the Smithsonian Institution’s Global Volcanism Program listed 1433 volcanoes with eruptions during the Holocene period (approximately the last 10,000 years) or exhibiting current unrest (http://volcano.si.edu/list_volcano_holocene.cfm). At the same time, WOVO listed 79 volcano observatories and related organizations in 33 countries, and the International Association of Volcanology and Chemistry of the Earth’s Interior (IAVCEI, <https://www.iavceivolcano.org/>) claimed more than 2000 scientists as members – numbers that signify a remarkable and welcome expansion of volcano research facilities worldwide since the pioneering Osservatorio Vesuviano was established more than 170 years ago.

The preceding paragraphs emphasize the importance of observatories in the development of modern volcano science because such

establishments foster *multidisciplinary investigations* fueled by *data*, two intertwined aspects of volcano studies that often are conceptualized separately as *research* and *monitoring*. In fact, nearly all volcano research relies on information collected at volcanoes, and nearly all volcano-monitoring data are useful for research. Observatories, whether real or virtual, place-bound or dispersed, are synergistic settings in which volcano science thrives. Data are an essential element, and data from four scientific disciplines are especially important, as described below.

Seismology, Geochemistry, Geology, and Geodesy: Four Pillars of Modern Volcano Science

Volcano Seismology

Seismology has been a cornerstone of volcano studies at least since Pliny the Elder noted the occurrence of earthquakes prior to the 79 CE eruption of Mount Vesuvius. Still today, earthquake activity is the most consistently reported indication of volcanic unrest. Seismic signals from volcanoes can be generated by several different processes, and they can be recorded at greater distances than volcano-related ground deformation or geochemical changes. For these reasons, seismology is an especially effective tool for exploring magmatic systems. Recent advances in digital seismic instrumentation with wide dynamic range and in sophisticated computational methods have enabled new insights into such topics as the configuration of magma reservoirs, migration of magmatic and hydrothermal fluids, and causes of volcanic tremor (e.g., Chouet 2003). Detailed studies of earthquake focal mechanisms enabled by such advances have uncovered evidence for magma-system pressurization (Moran 1994) and fluid migration (Foulger and others 2004) beneath volcanoes. Studies of repeating earthquakes (“drumbeats”) during dome-building eruptions have revealed that in some cases they are an integral part of the extrusion process and therefore can be diagnostic of an eruption’s end (Moran and others 2008; Thelan and others 2008). Although routine eruption prediction remains out

of reach, such discoveries hint at the possibility of anticipating both the beginning and end of eruptions in the not-too-distant future.

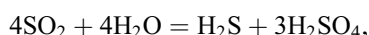
Gas Geochemistry

Hydrothermal activity and gas emission are two common occurrences at active volcanoes that can provide insights into subsurface processes. As is the case for seismology, the field of volcano geochemistry is too broad to be addressed in more than a cursory way here. Interested readers might want to consult, for example, a review of the subject by Wallace (2004). One topic in the field, volcanic gas flux measurements and their implications for subsurface magma volumes and eruption forecasts, is especially germane. Gas flux measurements and laboratory studies of solubility as a function of pressure support estimates of the volume and depth of degassing magma; time series flux measurements can be used to estimate a volcano’s magma supply rate (Gerlach and others 2002). Measurements of sulfur dioxide (SO₂) flux can be made with ultraviolet (UV) absorption spectrometers; fluxes of SO₂ and other gas species (HCl, HF, SO₂, CO₂, CO, and others) can be measured with a Fourier transform infrared spectrometer (FTIR) or by other means. A total flux measurement of one gas species (e.g., SO₂) can be combined with information about the chemical composition of the gas measured at a fumarole, for example, to estimate the fluxes of other species. See Sutton and Elias (2014) for a description of measurement techniques and results of volcanic gas studies at HVO throughout its hundred-year history.

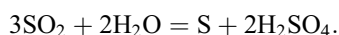
Gas flux measurements can be made from the ground or an aircraft in the vicinity of the volcano, or from Earth orbit with sensors such as the Moderate Resolution Imaging Spectroradiometer (MODIS) currently aboard two of NASA’s Earth Observing System satellites (<http://modis.gsfc.nasa.gov/>), the Ozone Monitoring Instrument (OMI) on the Aura satellite (<https://disc.gsfc.nasa.gov/Aura>), and the Ozone Mapping & Profiler Suite (OMPS) aboard the Suomi National Polar-orbiting Partnership (SNPP) spacecraft (<http://npp.gsfc.nasa.gov/omps.html>). Together these instruments provide daily global maps of

sulfur dioxide in the atmosphere (<http://so2.gsfc.nasa.gov/index.html>). For a discussion of spaceborne infrared techniques used to estimate properties and composition of volcanic ash clouds, including gases, see Pavolonis (2010) and Pavolonis and others (2013).

For eruption forecasting, more important than absolute gas fluxes are any changes in the relative emission rates of SO₂, H₂S, and CO₂. SO₂ is relatively soluble in water, so a groundwater system can oftentimes effectively “scrub” SO₂ exsolved from magma before it enters a volcanic plume. Scrubbing occurs as a result of two hydrolysis reactions (Symonds and others 2001):



and



At very “wet” volcanoes, groundwater scrubbing of SO₂ can be so effective that no precursory increase in SO₂ flux is observed before an eruption. A change from an H₂S-dominated plume to an SO₂-dominated plume can signal the arrival of new magma at shallow depth. Such a change can occur because (1) the kinetics of the first reaction cannot keep pace with an increased flux of SO₂ exsolved from newly arrived magma or (2) enough groundwater has been boiled off by magmatic heat to create dry passageways to the surface for SO₂. CO₂ is considerably less soluble in water than SO₂, so it mostly passes through the groundwater system and into the plume. The gas saturation depth in magma is much greater for CO₂ than SO₂, so CO₂ exsolves first and at greater depth. Consequently, a steady or increased CO₂ flux followed by an increased SO₂ flux can constrain models of deep magma supply followed by the arrival of fresh magma within a few kilometers of the surface.

Geology/Petrology

Too often considered to be a science of ancient rocks and landforms, geology also plays an important role in modern studies of active volcanism. The chemical composition and physical

state (crystallinity, volatile content, degree of alteration, etc.) of erupted products are directly linked to preeruptive conditions, and therefore they provide a valuable window into the subsurface. Systematic collection of samples throughout an eruption, especially when combined with rapid-turnaround analyses, put “petrologic monitoring” on par with geophysical, geochemical, and geodetic types of volcano studies. Resulting datasets enable insights into the evolution of magma storage and transport systems, including effects of fractionation, assimilation, and recharge (e.g., Thornber and others 2015). Repeated mapping of active flows and domes, whether in the field by conventional methods or remotely by aircraft or satellite, can be used to track effusion rate as a function of time, especially when volume estimates are obtained by photogrammetry or other means. Similar information can be gleaned from thermal surveys or direct measurements of lava flux (e.g., by measuring flow velocity in a lava tube of known cross-sectional area). Last but emphatically not least, the value of trained eyes in the field to make geological observations that might not be recorded by nonhuman sensors should not be underestimated. The human eye-brain combination has yet to be matched as a means to record and interpret natural phenomena.

Geodesy

Among the more extreme examples of volcano-related deformation is the case of Serapeo, a Roman marketplace constructed in the coastal town of Pozzuoli in the first century BCE. The structure has two generations of floors and columns: the second was built after the first sank below sea level. Borings in the columns by a marine mollusk indicate as much as 11 m of subsidence by about 1000 CE, followed by about 12 m of uplift from 1000 CE to 1538 CE. The uplift rate accelerated dramatically several years before an eruption that began on 29 September 1538. The most rapid uplift occurred on 26–27 September, when the ground level at Pozzuoli rose at least 4–5 m in 48 h, causing the shoreline to recede 400 m (Caputo 1979). The cause of the upheaval was magma intrusion beneath the partly submerged but still

active Campi Flegrei (Phlegraean Fields) caldera. Presumably unknown to its earliest inhabitants, Pozzuoli is situated near the caldera's center.

By measuring and modeling ground deformation at volcanoes, which in most cases is much more subtle than witnessed at Pozzuoli nearly 500 years ago, volcano scientists can infer the location, shape, and changing volume of magma reservoirs; formation and migration of dikes, sills, and eruptive conduits; accumulation and escape of magmatic volatiles; and pressure changes within hydrothermal systems (Dzurisin 2003, 2007). They have at their disposal an array of remote-sensing techniques and in situ sensors, notably the Global Navigation Satellite System (GNSS) and interferometric synthetic-aperture radar (InSAR), plus sophisticated numerical analysis tools (e.g., Segall 2010; McCaffrey 2009). The remainder of this article describes some of these techniques and how they are used to advance understanding of magmatic systems.

Techniques for Measuring Volcano Deformation

Classical Geodetic Surveying Techniques

Geodetic Leveling

Any technique for measuring precise locations of fixed points on Earth's surface (benchmarks) can be used to monitor volcano deformation by repeating surveys of a network of marks on and around the volcano. Classical surveying techniques including geodetic leveling, triangulation, and trilateration have been used at volcanoes to good effect for at least the past century. On the Big Island of Hawai'i, for example, the first leveling survey from the coast to the summit of Kīlauea Volcano was done in 1912; at the Yellowstone Caldera, Wyoming, the initial leveling survey was in 1923. In both cases, subsequent resurveys revealed vertical displacements caused by magmatic, tectonic, or landslide processes (e.g., Lipman and others 1985; Dzurisin and others 2012). Together with GPS, InSAR, and other geodetic tools, leveling surveys are still used by staff of the USGS Hawaiian Volcano Observatory as a

specialized tool to study specific deformation topics at Kīlauea Volcano. At Yellowstone and most other volcanoes, however, the time-honored technique has been supplanted by GPS and InSAR (see below).

Trilateration and Triangulation

Trilateration and triangulation are two other surveying techniques that were used widely during the twentieth century to study volcanoes but now have been mostly replaced by more modern methods (GPS, InSAR). By 1924, repeated triangulation and leveling surveys had documented uplift and extension across Kīlauea Caldera. Dr. Robert W. Decker organized the first trilateration surveys at Kīlauea and Mauna Loa volcanoes in the mid-1960s using electronic distance-measuring instruments (EDMs). Repeated surveys during ensuing decades helped to determine the location and size of primary and subsidiary magma reservoirs, and the volumes of magma injected into volcanic rift zones. Today, GPS and InSAR are the primary tools used for a wide variety of geodetic investigations at HVO and elsewhere around the world.

Global Positioning System (GPS/GNSS)

Fifty years ago, the US NAVSTAR Global Positioning System was a military secret; now space-based positioning pervades many aspects of everyday life. The US system was developed to provide global, 24-h, all-weather position and timing information primarily to military users. However, the popularity of its non-classified aspects quickly led to its widespread adoption and development by a large civilian community. An important breakthrough occurred when scientific users demonstrated that millimeter-scale positioning accuracy could be achieved by careful processing of the same signals that were intended to provide meter-scale accuracy for military users. In addition to NAVSTAR GPS, Russia operates its GLONASS satellite system, and the European Union's Galileo and China's BeiDou (COMPASS), both currently in deployment phase, are scheduled to be fully operational in 2020. The term GNSS (Global Navigation Satellite System) refers to all these systems; here we

use GPS to refer to precise positioning techniques that make use of the GNSS. For a discussion of GPS/GNSS principles and signal processing, see Dzurisin (2007, Chap. 4) and references therein.

For volcano studies, GPS is used primarily in three modes: two end-member modes known as campaign and continuous (Fig. 1) and a hybrid mode known as semicontinuous (SGPS) or semipermanent (SPGPS) (Hammond and others 2007; Blewitt and others 2009; Dzurisin and others 2017). The term “campaign” is a holdover from the days when GPS equipment was expensive, bulky, and power-hungry, so considerable planning and effort were required to observe a network of benchmarks for several hours to a few days each. As GPS equipment became more

affordable, power efficient, and smaller, more of it was installed at permanent sites and operated in continuous mode. Today, networks of permanent, continuously operating GPS (CGPS) stations are common. Campaign-style GPS is used mainly where strain rates are low, where CGPS stations are impractical (e.g., wilderness areas where permanent installations are not easily permitted), or to densify an existing network of CGPS stations. Hybrid SGPS or SPGPS, in which equipment is deployed for an extended period before retrieval, offers several advantages over campaign GPS, including better precision at lower cost (Dzurisin and others 2017). Generally, CGPS stations are the first choice because they provide an uninterrupted record and excellent resolution in



Volcano Deformation: Insights into Magmatic Systems, Fig. 1 Continuous GPS (CGPS) station AV37 at False Pass on Unimak Island, Alaska. Shishaldin volcano is visible in left background. Station is operated by UNAVCO as part of the Earthscope Plate Boundary Observatory. GPS antenna and enclosure (gray dome) are mounted atop a short-drilled, braced monument epoxied into rock to a depth of several feet (<http://www.unavco.org/instrumentation/monumentation/types/types.html>). GPS receiver, batteries, and telemetry equipment are in housing

at right with solar panels attached. View is to northeast. Inset: campaign-style GPS station near Mount St. Helens, Washington. GPS antenna is mounted atop a portable tripod and plumbed over a benchmark set in rock. Receiver and batteries are in yellow weatherproof housing. Data are recorded at site for retrieval after several hours to a few days. With a more stable antenna mount, such a setup can be operated in semicontinuous or semipermanent mode for up to several months where a more permanent CGPS station is impractical or unnecessary

the time domain (typically 15 or 30 s, 0.1–1 s possible). The latter characteristic is especially important during a rapidly developing volcanic event.

Two examples of extensive CGPS networks in countries with active volcanoes are the GPS Earth Observation Network (GEONET) in Japan and the Plate Boundary Observatory (PBO) in the western United States. GEONET was established in 1994 and comprises about 1230 CGPS stations with average spacing of 15–25 km throughout all of Japan (http://datahouse1.gsi.go.jp/terras/terras_english.html). The network provides detailed spatial and temporal information about ground deformation caused by earthquakes, volcanic activity, and collision of the Eurasian plate with the Pacific and Philippine plates. PBO consists of a network of about 1100 permanent CGPS stations, 78 borehole seismometers, 74 borehole strainmeters, 26 shallow borehole tiltmeters, and 6 long-baseline laser strainmeters (<http://pbo.unavco.org/>). The PBO CGPS network includes backbone stations with average spacing of 100–200 km that span the entire conterminous United States, plus several station clusters in the western United States with much tighter spacing: (1) the transform cluster in the vicinity of the San Andreas fault in California; (2) the subduction cluster along the Cascadia subduction zone in northern California, Oregon, Washington, and southern British Columbia; (3) the extension cluster in the Basin and Range region; and (4) the volcanic cluster at selected volcanoes in the Aleutian and Cascade arcs and at the Yellowstone and Long Valley calderas.

Interferometric Synthetic-Aperture Radar (InSAR)

What continuous GPS is to the time domain, InSAR is to the spatial domain. Classical geodetic surveys provide information only at benchmarks, which typically are 1–10s km apart. Leveling lines are located mostly along roads or other access routes that might not be well situated to characterize a deforming area, especially at a steep or remote volcano. Even dense CGPS networks have average station spacing of 10 km or more. While this might be adequate to characterize

broad-scale deformation, better spatial resolution is necessary to capture localized or complex deformation that sometimes occurs at volcanoes. With InSAR, it is possible to map an entire deformation field with a precision of ~ 1 cm and a spatial resolution of 1–10 m, and thus to obtain tighter constraints on any source(s) within the radar field of view (typical swath width for satellite-borne SARs is 10s–100s of km). Such fine spatial resolution comes at the price of poorer temporal resolution, however. Orbital repeat cycles for current SAR satellites can be as little as 1 day but are more generally in the range of a few days to weeks. As a result, the sampling frequency for ground deformation by repeat-pass InSAR can be no higher than 1 day, compared to 1 s or less for CGPS observations, for example. A useful strategy for volcano studies is to combine high-temporal-resolution CGPS data with high-spatial-resolution InSAR observations to infer an accurate space-time history for deformation. For additional information about InSAR and examples of what can be learned from complementary CGPS and InSAR studies, see Dzurisin (2007, Chap. 5), Lu and Dzurisin (2014), Pinel and others (2014), and the Case Studies section below.

As this is written in January 2019, six satellite missions suitable for InSAR studies are operational, and several are planned (Table 1). In the latter category is the NASA-ISRO Synthetic Aperture Radar (NISAR) mission, a joint project between the space agencies of the United States and India to launch a dual frequency (L-band and S-band) SAR satellite in 2022. The radars aboard past, present, and future SAR satellites operate at X-band (wavelength $\lambda \sim 3$ cm), C-band ($\lambda \sim 6$ cm), S-band ($\lambda \sim 12$ cm), or L-band ($\lambda \sim 24$ cm). Shorter wavelengths provide better ground resolution, but longer wavelengths achieve greater penetration through surface cover including vegetation and snow/ice. All the bands are useful for volcano studies.

The first satellite radar interferogram to grab widespread attention among the Earth science community appeared on the cover of the 8 July 1993 issue of the journal *Nature*. The image captured in dramatic fashion the displacement field produced by the 1992 magnitude 7.3 Landers

Volcano Deformation: Insights into Magmatic Systems, Table 1 Legacy and operational imaging radar satellites suitable for InSAR studies as of January 2019

Legacy				
Satellite	Operator	Launch	Frequency	Notes
Radarsat-1	Canada	1995	C-band	Non-operational as of 2013
ERS-1	European Space Agency	1991	C-band	Non-operational as of 2000
ERS-2	European Space Agency	1995	C-band	Non-operational as of 2011
JERS-1	Japan	1992	L-band	Non-operational as of 1998
ALOS-1	Japan	2006	L-band	Non-operational as of 2011
ENVISAT	European Space Agency	2002	C-band	Non-operational as of 2012
Operational				
Satellite	Operator	Launch	Frequency	Notes
COSMO-SkyMed	Italy	2007–2010	X-band	Four-satellite constellation with 1–16 day orbital repeat cycle for each satellite http://www.cosmo-skymed.it/en/index.htm
TerraSAR-X, TanDEM-X	Germany	2007, 2010	X-band	Twin-satellite constellation for global DEM generation https://www.dlr.de/dlr/en/desktopdefault.aspx/tabid-10376/
Sentinel-1A and sentinel-1B	European Space Agency	2014	C-band	Two-satellite constellation with combined six-day orbital repeat cycle http://www.esa.int/Our_Activities/Observing_the_Earth/Copernicus/Sentinel-1
Radarsat-2	Canada	2007	C-band	Follow-on to Radarsat-1. 1 m resolution in spotlight mode, 100 m in ScanSAR mode http://www.asc-csa.gc.ca/eng/satellites/radarsat2/default.asp
ALOS-2	Japan	2014	L-band	Follow-on to ALOS-1 with 1x3 meter pixels in spotlight mode, 10 m resolution in survey mode http://global.jaxa.jp/projects/sat/alos2/index.html
KOMPSAT-5	Korea	2013	X-band	http://www.si-imaging.com/products/

earthquake in California. The accompanying article by Massonnet and others (1993) opened many scientists' eyes to InSAR's potential for mapping ground deformation, and the first application to volcano deformation came shortly thereafter at Mount Etna (Massonnet and others 1995). The ensuing two decades have brought many refinements. Massonnet and Feigl (1998) introduced the idea of pairwise logic to distinguish between deformation on the ground, perturbations in the atmosphere, and artifacts in the processing of SAR data by comparing different interferometric pairs that span different time intervals. Biggs and

others (2007, 2010) enhanced the technique of stacking several InSAR images of the same area, which improves signal-to-noise, by adding the concept of chains to further reduce atmospheric artifacts. Perissin and Wang (2012) extended the chain concept to two dimensions, time and orbital separation. Feigl and Thurber (2009) described a strategy for choosing the optimum set of pairs, and Agram and others (2013) developed a software toolbox for InSAR time series analysis that incorporates several of these refinements.

As additional SAR satellites become operational and automated data processing techniques

are developed, the production of standard, two-pass interferograms is becoming routine. Techniques such as persistent scatterer InSAR (PS-InSAR) and small baseline subset InSAR (SBAS) have been developed to address specific problems and thus to extend the utility of InSAR in general, including its application to volcano deformation (Ferretti and others 2000, 2001, 2011; Bonforte and others 2004; Salvi and others 2004; Hooper and others 2004, 2007; Perissin and Wang 2012; Simons and others 2011; Hetland and others 2012; Agram and others 2013; Lu and Dzurisin 2014, Chap. 3). In the future, deciphering time-dependent deformation signals will require hybrid algorithms that combine the distinct advantages of each technique into a more holistic approach (Hooper 2008; Hooper and Zebker 2007; Hooper and others 2004; Hole and others 2006; Agram and others 2013; Pinel and others 2014).

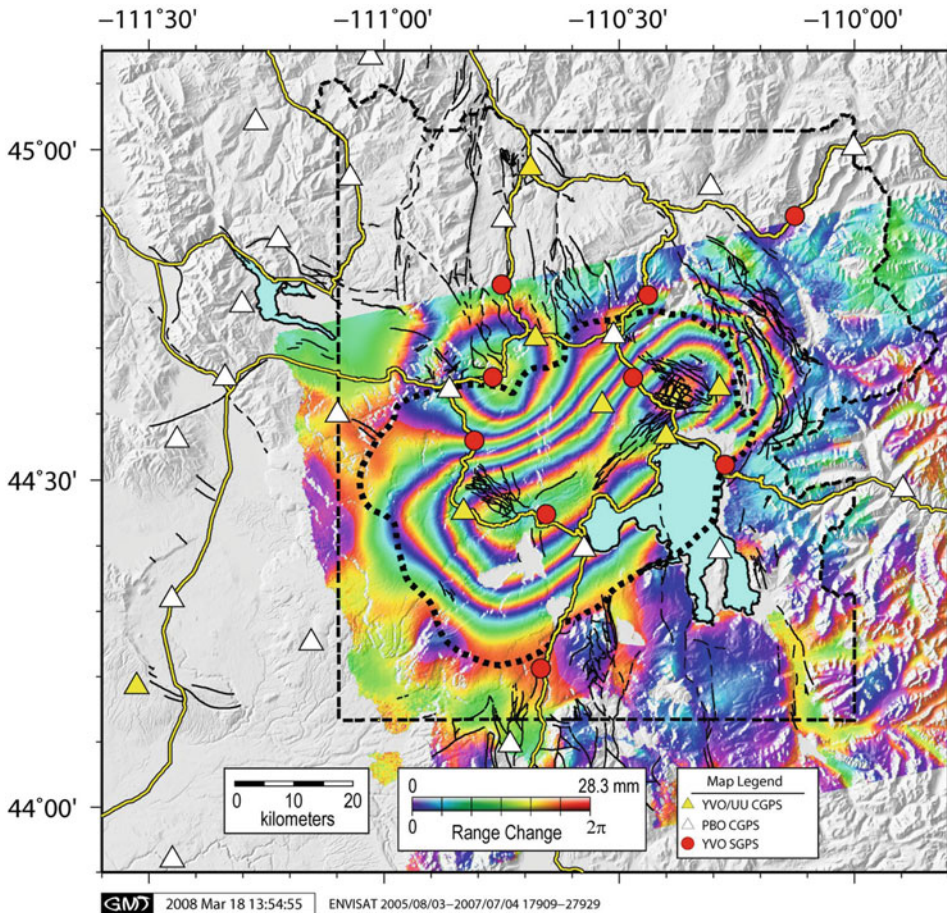
Until recently, most InSAR analyses have been opportunistic, i.e., reliant on data acquisitions by SAR satellites with disparate designs and mission objectives. To address this issue, the US National Aeronautics and Space Administration (NASA) and the Indian Space Research Organization (ISRO) have proposed the NISAR mission specifically to address complex Earth processes, including ecosystem disturbances, ice-sheet collapse, and natural hazards such as earthquakes, tsunamis, volcanic eruptions, and landslides (<http://nisar.jpl.nasa.gov/nisarmission/>). Concurrently, the Advanced Rapid Imaging and Analysis (ARIA) Center for Natural Hazards at the Jet Propulsion Laboratory, California Institute of Technology, is developing automated geodetic imaging and analysis capabilities to combine information from InSAR, GPS, and other space-based techniques into an operational tool in support of hazards-response communities. Primary goals of the ARIA Center are to (1) generate imaging products in near-real time to improve situational awareness for disaster response and (2) provide automated imaging and analysis capabilities for researchers to take advantage of the dramatic increase in data expected from planned imaging radar satellite missions (Table 1) (<http://aria.jpl.nasa.gov/about>).

The main advantage of InSAR over GPS and other geodetic monitoring techniques is illustrated in Fig. 2, which shows a deformation interferogram for the time interval from 3 August 2005 to 4 July 2007 at Yellowstone National Park, Wyoming. The pattern of repeating color bands represents interferometric fringes produced by uplift of the caldera floor at an average rate of ~75 mm/year and simultaneous subsidence of an area along the north caldera rim at an average rate of ~35 mm/year. Also shown in the figure are the locations of CGPS and SGPS stations in the region. Even with this relatively dense network of GPS stations, the interferogram provides a much clearer view of the spatial pattern of ground deformation. The GPS stations, on the other hand, provide much better temporal resolution: Raw data are acquired at 1–15 s intervals and analyzed as daily estimates of relative position.

Lidar

A half-century after the first EDM surveys at Kīlauea and Mauna Loa, precise distance measurement with lasers has morphed into lidar. The term “lidar” commonly is described as an acronym of *light detection and ranging*. However, the first documented use of the term was as a portmanteau (blend) of “light” and “radar” (Ring 1963). Lidar is a revolutionary mapping tool with applications to volcanoes worldwide. Gone is the necessity for ground access to a field area with bulky tripods, EDM, and manufactured reflectors to be painstakingly set up over a network of benchmarks. Enter lidar, the laser mapping tool that can be used from the ground in a mode akin to a twenty-first century EDM (Collins and Kayen 2006) but also can be fitted to aircraft and satellites to scan large swaths of Earth’s surface, including volcanoes.

Lidar uses a pulsed laser to precisely measure distances from the instrument to a target, commonly Earth’s surface. Most lidars operate in the near-infrared portion of the electromagnetic spectrum at wavelengths of 600–1000 nm. Airborne topographic mapping lidars generally use yttrium aluminum garnet (YAG) lasers that operate at 1064 nm with pulse repetition rates of 50–100 kHz. When the reflected signals are



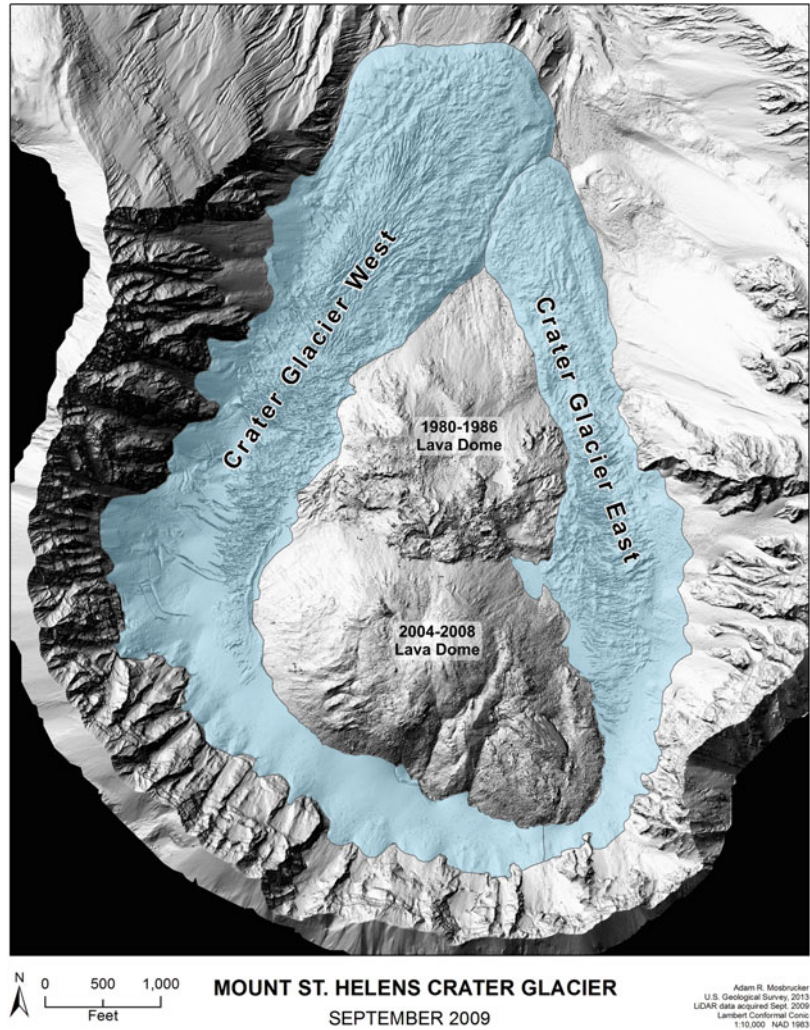
Volcano Deformation: Insights into Magmatic Systems, Fig. 2 Deformation interferogram produced by C. Wicks (USGS) from Envisat satellite images acquired on 3 August 2005 and 4 July 2007 showing uplift of the 640,000-year-old Yellowstone Caldera (heavy dotted line) and subsidence of a smaller area along the north caldera rim centered near Norris Geyser Basin. Each interferometric fringe, represented by a range of colors from violet to green to yellow to red or vice versa, represents 28.3 mm of range change (mostly uplift or subsidence in this case) between the ground and satellite. The central part of the caldera rose at an average rate of ~75 mm/year, while the

Norris area subsided ~35 mm/year. Also shown are continuous and semicontinuous (semipermanent) GPS stations in the vicinity of Yellowstone National Park (dashed line, with major roads shown by solid yellow lines). Yellow triangles, CGPS stations operated by the Yellowstone Volcano Observatory (YVO) and University of Utah (UU). White triangles, CGPS stations operated by the Plate Boundary Observatory (PBO). Red circles, SGPS (SPGPS) stations deployed by YVO during summer only, first in 2008 (6 stations) or 2009 (3 additional stations) and yearly since then

processed together with position and orientation data, scan angles, and calibration data, the result is a dense cloud of points, each with three-dimensional spatial coordinates. Vertical accuracy on the order of centimeters is achievable with spatial resolution on the order of meters (Fig. 3). Lidar's high sampling frequency (pulse repetition rate) supports the grouping of return signals in

multiple layers of information about the target. First returns from each resolution cell are associated with the highest features in the landscape (typically treetops, tops of buildings, etc.), and subsequent returns generally are associated with understory vegetation. The final return from each cell is associated with the bare Earth, i.e., it results from a pulse that penetrated any vegetation

Volcano Deformation: Insights into Magmatic Systems, Fig. 3 Hillshade derived from lidar data showing summit area and 1980 crater at Mount St. Helens, Washington. Horseshoe-shaped crater rim surrounds Crater Glacier, which formed after the catastrophic eruption of May 18, 1980. Glacier in turn surrounds lava domes that grew from 1980 to 1986 and from 2004 to 2008. Lidar data acquired in 2009 and available in USGS Data Series 904 (Mosbrucker 2014)

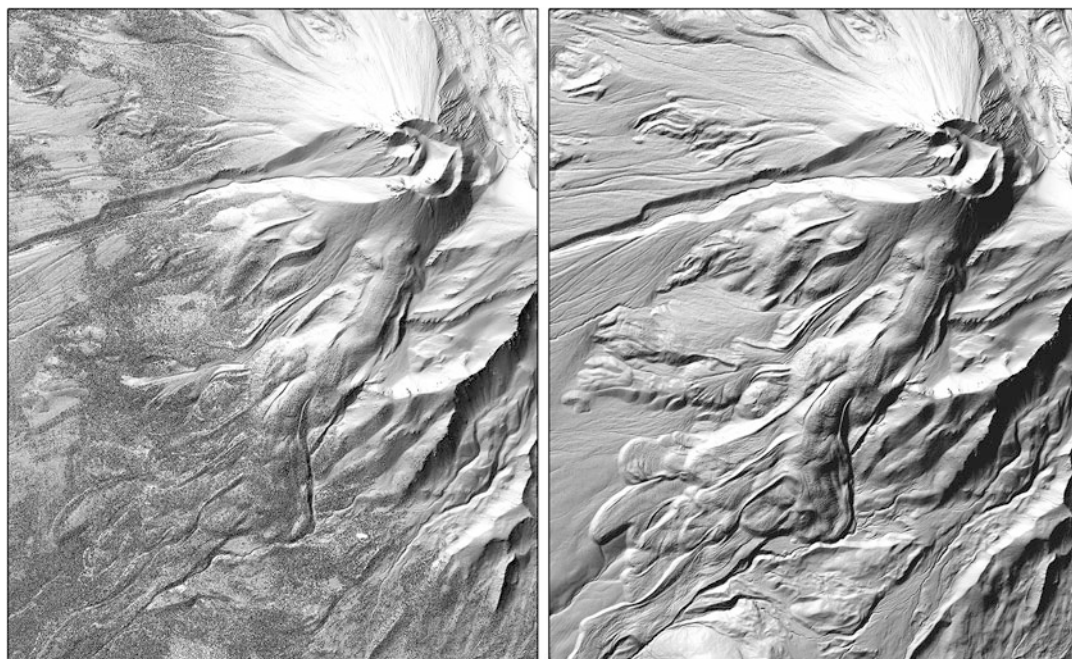


canopy to reach the ground surface and return. By using only the latest returns, a bare-Earth digital elevation model (DEM) can be created, together with derived products such as hillshades and inundation maps (Fig. 4).

Bathymetric lidars operate in the green portion of the spectrum to penetrate the water column and map the seafloor, lakebeds, and river beds. An example is the USGS Experimental Advanced Airborne Research Lidar (EAARL), an airborne bathymetric lidar system for surveying coral reefs, nearshore benthic habitats, coastal vegetation, and sandy beaches (Wright and Brock 2002).

Lidar can be used as a tool for mapping ground deformation by differencing sequential lidar-

derived DEMs of a deforming area. Centimeter-level accuracy is achievable from ground-based lidar surveys under favorable conditions, i.e., in the absence of rugged terrain or dense vegetation, and decimeter-level accuracy can be obtained from airborne surveys (Jaboyedoff and others 2012). In some cases, hazards near a restless volcano preclude repeated ground-based lidar surveys. However, the technique has been used to map inflation of active pāhoehoe flows at Kīlauea Volcano, Hawai‘i, for example. Airborne lidar surveys were used to track the evolution of an active flow field at Mount Etna, Italy (Favalli and others 2010), and would be adequate to monitor deformation caused by cryptodome growth, for example.



Volcano Deformation: Insights into Magmatic Systems, Fig. 4 Hillshades derived from lidar data showing Shastina cone on the west flank of Mount Shasta, California. Left image shows topography plus vegetation (trees);

right image shows bare Earth with vegetation removed. Lidar acquired in 2010 and available in USGS Data Series 852 (Robinson 2014)

Photogrammetry

Classical photogrammetric techniques were more often used to produce topographic maps than to monitor active ground deformation but that has changed with the development of new tools that enable rapid production of DEMs from photographs acquired with low-cost digital cameras using commercially available software (e.g., Diefenbach and others 2012; Thompson and Schilling 2007). Diefenbach and others (2012) demonstrated that oblique photos taken from a helicopter using a digital single-lens reflex (DSLR) camera could be used to produce sequential DEMs of a dacite lava dome that grew at Mount St. Helens from 2004 to 2008 and that associated volume estimates consistently agreed with those from traditional vertical photogrammetric surveys to within 5%. Today, with access to inexpensive software, a DSLR, and a handheld GPS unit, users can produce DEMs in a matter of hours to characterize surface features and map changes in a rapidly evolving landscape.

Tralli and others (2005) discussed the role of remote-sensing techniques, including lidar and photogrammetry, in advancing understanding of processes that lead to natural hazards (e.g., earthquakes, volcanic eruptions, floods, landslides, and coastal inundation). With regard to volcano hazards, they wrote (p. 185): “Seismic and in situ monitoring, geodetic measurements, high-resolution digital elevation models (e.g., from InSAR, Lidar and digital photogrammetry) and imaging spectroscopy (e.g., using ASTER, MODIS and Hyperion) are contributing significantly to volcanic hazard risk assessment, with the potential to aid land use planning in developing countries where the impact of volcanic hazards to populations and lifelines is continually increasing.” The remainder of this section deals with tiltmeters, strainmeters, and microgravity surveys that, together the space-based methods discussed above, can be used to investigate subsurface processes at volcanoes and to assess the associated hazards.

Tiltmeters and Strainmeters

Whereas GPS provides three-dimensional positioning information and therefore can be used to measure 3D surface displacements, other types of sensors track different characteristics of the deformation field that can be useful at volcanoes. Tiltmeters measure very small changes in surface inclination that can be caused, for example, by subsurface changes in volume or pressure, or by fluid movement. Modern electronic tiltmeters use electrodes to sense the exact position of a bubble in an electrolytic solution, which changes when the sensor is tilted. The tiltmeter housing can be mounted to the ground either horizontally (platform type) or vertically in a borehole up to several hundred meters deep (borehole type). Sophisticated tiltmeters are capable of detecting tilts (or changes in topographic slope) as small as 5 nanoradians, which corresponds to a vertical rise or fall of about 5 micrometers (5×10^{-6} m) over a horizontal distance of 1 km. Greater sensitivity is the primary advantage of tiltmeters over CGPS and most other deformation sensors or geodetic techniques; tiltmeter data are also easier to process than GPS data, although that advantage has largely disappeared as rapid GPS data processing has been automated on modern computing systems. Like CGPS, tiltmeters provide excellent temporal resolution for deformation signals, but spatial resolution is limited by the density of stations in the sensor network. Near-surface tiltmeter installations can be plagued by site-specific or extraneous effects such as temperature, ground-water fluctuations, or long-term drift. A key factor in borehole installations, whether of tiltmeters or strainmeters (see below), is the quality of coupling between the sensor and surrounding rock. The effects of imperfect coupling and external influences such as ocean tides, solid Earth tides, and pore-pressure variations can be modeled empirically and then removed to improve the utility of the data for deformation modeling (e.g., Canitano and others 2014).

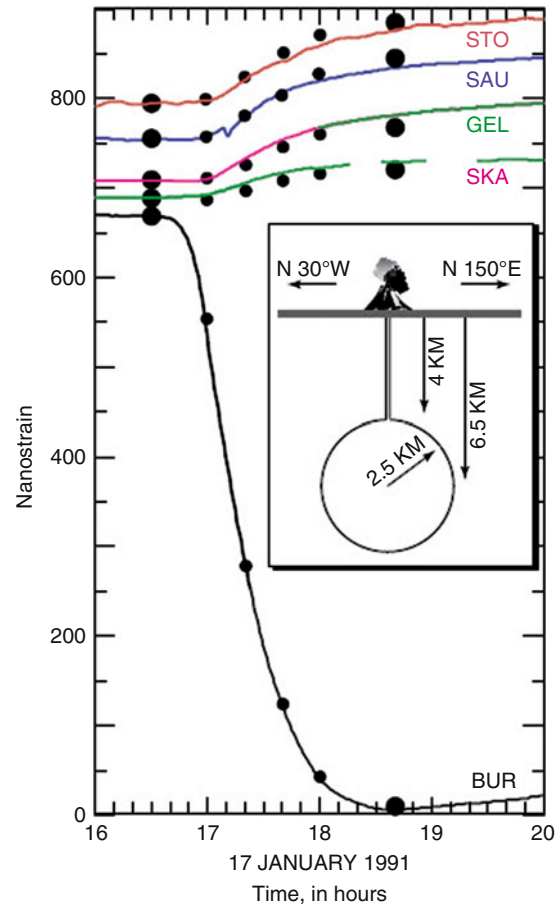
Strainmeters measure changes in the shape or volume of a body as a result of stress. There are many designs capable of measuring linear, areal, volumetric, or shear strain, but the most common types used for volcano monitoring are the Sacks-

Evertson dilatometer and the Gladwin tensor strainmeter. Both types typically are installed in boreholes 100–250 m deep to isolate them from surface noise. The Sacks-Evertson dilatometer consists of a stainless-steel cylinder with an annulus filled with silicone oil. As the cylinder is deformed, oil is forced in or out of an attached bellows. The top surface of the bellows consequently moves up or down by an amount proportional to the annulus volume divided by the cross-sectional area of the bellows. That ratio is large ($\sim 4 \times 10^4$), resulting in a sensitivity of $\sim 10^{-12}$ in volumetric strain. A three-component design operates on the same hydraulic principle but incorporates three sensing volumes oriented at 120° with respect to each other, enabling the determination of shear strain as well as areal or volumetric strain. The Gladwin tensor strainmeter incorporates four strain gauges (one is a spare), likewise oriented 120° apart, that can detect changes in the nominal 10-cm diameter of the strainmeter on the order of 4×10^{-10} cm, which equates to approximately 0.05 nanostrain (5×10^{-11}). For comparison, crustal strains induced by solid Earth tides are tens of nanostrain in magnitude, and much larger strain signals typically accompany volcanic activity. Gladwin tensor strainmeters were installed at several sites as part of the Plate Boundary Observatory (PBO), including small networks at Mount St. Helens, the Yellowstone Caldera, and several volcanoes along the Aleutian arc.

A notable success in the use of strainmeters for eruption prediction occurred at Hekla volcano, Iceland, in February 2000. A network of 5 Sacks-Evertson dilatometers was installed in 1979 to monitor strain changes associated with earthquakes in the South Iceland Seismic Zone. Coincidentally, one of the instruments was located 15 km from Hekla volcano; the others were 35–45 km away. In spite of the considerable distances involved, all 5 dilatometers recorded strain changes during an eruption of Hekla in January 1991 (Fig. 5). More importantly, the closest instrument started recording changes 30 min *before* the eruption began (Linde and others 1993). Nine years later, recognition of a remarkably similar pattern of local earthquakes and strain

Volcano Deformation: Insights into Magmatic Systems, Fig. 5

Borehole strain data for a 4-h period that includes the onset of the 1991 eruption of Hekla Volcano, Iceland. Inset shows model used by Linde and others (1993) to calculate points overlaid on the curves. The exact onset time of the eruption is not known from direct observations owing to inclement weather. However, airport radar observations set the time between 17:00 and 17:10 h (when the plume was already 11.5 km high), and farmers near the volcano reported seeing activity at ~17:05 h. All times are local. (Figure reproduced from Dzurisin 2007, p. 320)



changes at Hekla enabled a public warning to be broadcast about 20 min before the start of another eruption (Agustsson and others 2000; Linde and Sacks 2000). Since then, Sacks-Evertson borehole strainmeters have been installed near the Soufriere Hills volcano, Montserrat; Long Valley-Mono-Inyo volcanic chain, California; Kilauea and Mauna Loa volcanoes, Hawai'i; and Vesuvius, Campi Flegrei, Stromboli, and Mount Etna, Italy.

Microgravity Surveys

Studies of ground deformation, regardless of how detailed or accurate they might be, cannot answer one fundamental question about noneruptive activity at a volcano: Has the amount of magma at shallow depth beneath the volcano changed? This is because magma accumulation or movement is only one of several possible causes of volcano deformation. Others include the

accumulation or loss of magmatic volatiles and pressurization or depressurization of a hydrothermal system. Both processes can cause surface deformation, but they differ from magma accumulation/movement is one important way: Changes in magma storage are accompanied by significant changes in subsurface mass distribution, whereas the other processes are not. Microgravity surveys are uniquely capable of detecting mass redistribution, hence their importance for volcano studies.

The local gravitational acceleration g at any point on Earth's surface is given by the differential form of Gauss's law: $\nabla \cdot \mathbf{g} = -4\pi G\rho$, where ∇ denotes divergence, G is the universal gravitation constant, and ρ is the mass density at each point. If the spatial distribution of density ρ is spherically symmetric, then the equation reduces to the more familiar form: $g = \frac{GM}{r^2}$, where M is the mass of

Earth and r is distance from Earth's center. In the real-world case that Earth's density distribution is not spherically symmetric, the value of g at any point on the surface depends on both the mass distribution beneath the point and on the point's distance from Earth's center (i.e., elevation). Those dependencies provide a means to differentiate between two very different scenarios at an inflating volcano. In the first scenario, an intrusion of magma beneath the volcano results in both surface uplift and a changed subsurface mass distribution. In the second, pressurization of a magma storage zone or hydrothermal system causes uplift, but no appreciable change in mass distribution. Simultaneous, collocated elevation and gravity surveys before and after the inflation event can be used to differentiate between the two scenarios. The effects on g of any measured height changes (from leveling or GPS surveys, for example) can be removed by making a free-air correction of $\partial g/\partial z \cdot \Delta h$, where $\partial g/\partial z$ is the vertical gravity gradient in mGal/m and Δh is the height change in meters. $1 \text{ mGal} = 10^{-3} \text{ Gal} = 10^{-5} \text{ m/s}^2$. The so-called "normal" free-air gravity gradient, $\partial g/\partial z = -0.3086 \text{ mGal/m}$, is widely used to remove the effect of height changes from repeated gravity surveys. Measurements show that the actual value of $\partial g/\partial z$ varies by up to 20% depending on location, geology, and local topography (Hunt and others 2002). In most cases, the normal gradient is adequate to correct observed gravity changes to within the uncertainty of the measurements. Any residual change in g between the two surveys can be attributed to a change in subsurface mass distribution, which would be positive in the first case (mass increase) and near zero in the second.

Brown and Rymer (1991) reviewed the use of microgravity surveys to monitor active volcanoes and discussed implications of the observed gravity gradient, $\Delta g/\Delta h$, for subsurface processes. For example, in the case that the value of $\Delta g/\Delta h$ exceeds that of the free-air gradient, possible causes include magma intrusion, bubble collapse (densification), or a rise in the water table. Dzurisin and others (1980) and Johnson and others (2010) reported such an occurrence at Kīlauea Volcano, Hawai'i, following the

29 November 1975 M 7.2 earthquake beneath the volcano's south flank. For two overlapping time periods (December 1975–April 1977 and November 1975–January 2008), the observed residual gravity gradient exceeded the free-air gradient, which led the authors to conclude that the changes were caused by refilling of void space created when magma withdrew from beneath the summit in response to the quake.

Repeated gravity surveys have produced notable results at diverse volcanoes (e.g., Smith and others 1989; Arnet and others 1997; Battaglia and others 2003; Dzurisin and others 1980; Furuya and others 2003; Jachens and Eaton 1980; Johnson 1995; Jousset and others 2000; Rymer and others 1995; Vasco and others 1990; Williams-Jones and Rymer 2002; Carbone and others 2003a, b; Kauahikaua and Miklius 2003; de Zeeuw-van Dalfsen and others 2005; Berrino and others 2006; Bonafede and Ferrari 2009), but their use has not been widespread, due in part to the high cost of gravity meters suitable to the task (~US\$100 K).

As is the case for GPS, gravity meters can be used in either campaign or continuous mode. Both uses have produced notable results at Kīlauea Volcano, Hawai'i, and at Mount Etna, Italy (Rymer and others 1995; Carbone and others 2003a, b). At Kīlauea, fluctuations in continuous gravity records from sites near the summit have been attributed to convection in a shallow magma reservoir (Carbone and Poland 2012). Modeling of an abrupt change in gravity associated with draining of a summit lava lake produced the surprising result that the density of the lake was $950 + 300 \text{ kg/m}^3$ – comparable to the density of water and more than 50% less than the density typically assumed for basaltic magma – implying that the lake is very gas-rich (Carbone and others 2013). Campaign-style gravity surveys over a 3-year period, corrected for surface height changes measured with GPS and InSAR, showed net mass accumulation in the same area – in this case attributed to replacement of low-density, gas-rich magma in the reservoir by denser, degassed magma from above (Bagnardi and others 2014). Earlier, Johnson and others (2010) attributed progressive mass accumulation that was

detected by microgravity surveys from 1975 to 2008 to the accumulation of magma in a network of interconnected fractures, possibly created when magma withdrew from beneath the summit area in response to the 29 November 1975 *M* 7.2 earthquake beneath the volcano's south flank.

Modeling and Analysis Techniques

Data acquisition is a primary step in most volcano studies (other than purely theoretical investigations, which are valuable in their own right), but the next step – analysis and interpretation – almost always involves some form of modeling. This can be qualitative, such as a conceptual model to explain the rapidly receding shoreline at Pozzuoli in 1538 as a consequence of ground uplift caused by magma intrusion. On the other hand, sophisticated quantitative models that are constrained by multiple datasets and treat uncertainties in a rigorous way are becoming the norm. What follows is a brief description of various deformation source models and modeling techniques in common use today. Interested readers will find a thorough treatment of earthquake and volcano deformation in Segall (2010) and a more focused application to volcanoes in Lisowski (2007).

The goal of deformation source modeling is to infer the locations and characteristics (size, shape, strength, etc.) of subsurface elements responsible for an observed surface deformation field. For example, from measurements of the pattern of surface uplift at a volcano, we might attempt to constrain the subsurface boundaries, the idealized shape of those boundaries (sphere, dike, sill, pipe), and dilation of that (presumed) source of deformation, which plausibly might be a subsurface magma reservoir. To simplify this problem, we need to make assumptions about the mechanical properties of the crust beneath the volcano, which in reality might be quite complex.

Volcano deformation modeling is an inherently nonunique problem, despite simplifying assumptions often made to model deformation data. Given a goodness-of-fit criterion and a sufficient number of free model parameters, an observed deformation field can almost always be fit by

any combination of sources. For example, uplift from various axisymmetric sources (cylinder, dike, or sill) can be made identical to that from a sphere simply by varying their depth. They can be distinguished only by considering the horizontal components of the surface displacement field. For a modeling effort to be robust, ancillary data must be included in a rigorous way.

Considerable progress has been made recently toward developing volcano source models that include some of the physics of the causative processes, with associated uncertainties, rather than relying solely on sterile mathematical formulations or pattern recognition. In a review of eruption forecasting methods, Sparks (2003, p. 1) wrote: "The coupling of highly non-linear and complex kinetic and dynamic processes leads to a rich range of behaviours. Due to intrinsic uncertainties and the complexity of non-linear systems, precise prediction is usually not achievable. Forecasts of eruptions and hazards need to be expressed in probabilistic terms that take account of uncertainties." Deformation data and deformation processes are only one component of most physics-based volcano models, but together with their counterparts from seismology, geochemistry, and petrology, they provide a basis for quantitative forecasts of eruptions and associated hazards. Further discussion of physics-based volcano models is beyond the scope of this article, but for those interested in this topic, articles by Sparks (2003) and by Anderson and Segall (2011) are good sources of additional information.

Analysis of geodetic and geophysical data can rely on forward models for guidance or can proceed through inference using inverse methods. Forward models use physical principles and known or assumed material properties to calculate a theoretical system response. Forward models are useful to gain some understanding of a system, particularly if the physical system is understood well enough that the initial and boundary conditions are known. In this case, equations governing the system response can be formulated and solved. More commonly, we lack critical data on an exceedingly complex system (such as an erupting volcano), and we must infer system conditions and material properties through inverse

methods. Usually it is impossible to analyze geophysical data without making some inferences and using inverse methods. Typically what we have available is a generalized forward model, with initial and boundary conditions, and some observational data. Through systematic comparison, the best fitting parameters for the model can be determined with some uncertainty, but the sole use of the best fitting result, the maximum likelihood result, often leads to over-fitting or grossly underestimating the uncertainty in the solution. One way this limitation is overcome is by using Bayesian inference, which delivers a probability distribution of results that includes but is not limited to the maximum likelihood solution.

Analytical Deformation Source Models

Analytical deformation source models are mathematical formulations that describe surface deformation produced by subsurface forces or displacements, and so in practice they require simplifying assumptions about the behavior of the real Earth. A common assumption is that the deformation source is embedded in a half-space (an infinite domain bounded by a horizontal and planar surface free of any traction) which has uniform and isotropic elastic material properties. Although this is clearly an oversimplification, empirically this theoretical geometry produces solutions that compare well with reality, except near strong crustal heterogeneities or in areas of steep topography. Of course, the flanks of many volcanoes are steep, and their underpinnings are both heterogeneous and inelastic (hot). Corrections for topography and Earth curvature can be applied, and elastic heterogeneity can be included in the modeling process. So, too, can viscoelastic and poroelastic effects if the observational constraints are sufficiently strong to justify such an approach (Segall 2010).

Several analytical source models have been developed specifically for, or are directly applicable to, volcano deformation problems. No discussion of the topic would be complete without a nod to the time-honored “Mogi model” popularized by Japanese seismologist Professor Kiyoo Mogi in 1958. Mogi (1958) attributed the model’s mathematical formulation to Yamakawa (1955),

although it had been independently derived by both Anderson (1936) and McCann and Wilts (1951). Nonetheless, Mogi’s application of the model to two iconic eruptions – the great 1914 eruption of Sakurajima volcano, Japan, and the 1924 phreatomagmatic eruption of Kīlauea Volcano, Hawai‘i – earned him an enduring place in the annals of volcanology. Today, it is a rare publication dealing with volcano deformation that does not reference the well-known and still very useful “Mogi model.” To represent a magma reservoir, his formulation envisions a spherical pressure source (cavity) of radius α embedded at depth d in an elastic half-space such that $\alpha \gg d$. In the limit that $\frac{\alpha}{d} \rightarrow 0$, we can speak of a “point pressure source” or simply “point source”; both are common shorthand for the Mogi formulation. The surface displacements produced by a pressure change ΔP in a Mogi source are given by:

$$\begin{pmatrix} u \\ v \\ w \end{pmatrix} = \alpha^3 \Delta P \frac{(1 - \nu)}{G} \begin{pmatrix} \frac{x}{R^3} \\ \frac{y}{R^3} \\ \frac{z}{R^3} \end{pmatrix}$$

where u , v , w are displacements on the surface at the point $(x, y, 0)$, the center of the cavity is at $(0, 0, -d)$, ν is Poisson’s ratio which is often taken to be 0.25, G is the shear modulus, and $R = \sqrt{x^2 + y^2 + z^2}$ is the radial distance from the center of the cavity to a point on the free surface. The net surface displacement vector U_r is radial to the source, and its magnitude varies with the inverse square of the distance from the source:

$$U_r = \sqrt{u^2 + v^2 + w^2} = \alpha^3 \Delta P \frac{1 - \nu}{G} \frac{1}{R^2}.$$

In spite of simplifying assumptions that are not strictly satisfied in the real Earth, the Mogi model very often does an adequate job of fitting volcano deformation data, except for sources that are unusually shallow or elongate (e.g., feeder conduits, dikes, sills). For the case of a finite spherical cavity, i.e., $\frac{\alpha}{d}$ is not negligibly small, McTigue (1987) derived higher-order expressions for

surface displacements caused by a pressure change at the source. To a first approximation, the maximum displacements for a finite sphere are about $1 + (\frac{x}{a})^3$ times those for a point sphere, but the patterns of deformation are very similar (Lisowski 2007, p. 290), as is the vertical displacement field for any axisymmetric source whose vertical axis is orthogonal to the planar free surface (Dieterich and Decker 1975).

Spherical deformation sources, whether Mogi-type or finite, generally are not appropriate to represent magma feeder conduits; a pipe or elongate cylinder is more realistic. Bonaccorso and Davis (1999) formulated a model for a closed vertical pipe embedded in an elastic half-space. A closed pipe generates proportionally more horizontal displacement than a point source or finite spherical source, the vertical and horizontal displacements decay more slowly, and a profile of vertical displacement has a characteristic dimple over the pipe (Lisowski 2007, p. 292) (see Fig. 4 in Dzurisin 2003).

Various analytical source models have been formulated to approximate the effects of dike or sill-like magma intrusions or reservoirs. These include finite rectangular tensile dislocations (Davis 1983; Yang and Davis 1986), pressurized triaxial ellipsoids (Davis 1986), and finite pressurized horizontal circular cracks (Fialko and others 2001). The most appropriate formulation to use for a given modeling problem depends on what is known about the subsurface from other disciplines (e.g., geology, seismology, geophysics) and on the details of the deformation datasets to be modeled (i.e., spatial resolution, accuracy, complexity). For additional information, interested readers are encouraged to consult the original literature cited above or detailed treatments of the topic by Lisowski (2007) or Segall (2010).

In the case that little is known about the likely shape of a deformation source, two or more source models can be fit to observations and their results compared to determine the best choice. The principle of Occam's razor, which states that among competing hypotheses (models) the one with the fewest assumptions should be selected, is commonly applied. Stated differently, given two models that satisfy equally all available

observational constraints, the model with the least number of adjustable parameters is preferred. More complex models simply are not justified (cannot be constrained) with available observational data. To apply this principle, volcano modelers can use several methods of statistical inference. Bayesian methods are being used more often in recent years and automatically incorporate Occam's razor. As new observations become available, they are tested to see if they supply original constraints on the models.

Numerical Deformation Source Models

Analytical solutions cannot be found for the deformation produced by any but the simplest sources. Numerical techniques allow for approximate solutions to more complex problems – for instance, the deformation produced by an arbitrarily shaped source in a complex medium with surface topography. The precision and resolution of modern geodetic techniques such as InSAR, GPS, and repeated gravimetric surveys are in some cases sufficient to justify such a modeling approach. The finite element method (FEM) is a technique for solving the partial differential equations for conservation of mass and momentum that govern a medium such as the rock surrounding a volcano. The method has proven useful for computing deformation without neglecting such complexities as irregular source geometries, topography, heterogeneous material properties, and differing crustal rheologies (elastic, viscoelastic, viscous, plastic) – all of which undoubtedly exist in active magmatic systems.

In FEM analysis, the spatial domain (such as the medium surrounding a magma reservoir) is subdivided into a large number of simpler parts, called finite elements, in order to find an approximate solution to an otherwise intractable problem. Software packages are available for designing finite element meshes, solving the resulting equations, and displaying results. FEM techniques are much more computationally demanding than analytical modeling techniques, but this limitation is not important in many cases given dramatic increases in computing power for widely available desktop systems. FEM analysis has been applied to a wide variety of problems in

volcanology (e.g., Trasatti and others 2003; Del Negro and others 2009; Currenti and others 2010; Bonaccorso and others 2013) and is becoming more widely accessible to nonspecialists (Hickey and Gottsmann 2014). For a detailed treatment of the method and its applications, the reader might want to consult a textbook on the subject such as Reddy (2005) or research articles by Masterlark (2007), Masterlark and others (2010, 2011, 2012), Pedersen and others (2009), Ronchin and others (2013), and Ali and others (2014).

Two other numerical modeling techniques that have been applied to volcano problems are the finite difference method (FDM) and the boundary element method (BEM). As is the case for the FEM, these techniques are used to solve partial differential equations that govern the behavior of rock under stress. In the FDM, the differential equations are mapped onto a prismatic domain as difference equations written in terms of points used to define the domain. Finite differences in dependent variables between points are used to approximate the derivatives in the governing equations. In general, FDM is easier to implement than FEM but cannot handle complicated geometries easily. In the BEM, integrals of the conservation equations of interest are constrained with appropriate boundary conditions that relate the boundary solutions to the solution at interior points within the domain. BEM is less widely used than FEM or FDM but combines the efficiency of analytical modeling with the ability to include complex distributions of elastic material properties and geometries. Numerical methods in Earth science are the subject of books by Turcotte and Schubert (2002) and Gerya (2010); interested readers will find derivations, exercises, and useful software codes there.

The Inverse Problem in Volcano Geodesy

Given one or more deformation datasets and an appropriate model, how can we determine values for model parameters that best fit the observational data? This is an inverse problem, in which one attempts to infer the properties of a volcanic source of deformation from a set of measurements (e.g., the location, shape, and other physical characteristics of a subsurface deformation source).

A general discussion of inversion techniques in volcano geodesy is beyond the scope of this article. Earnest readers who are mathematically inclined might want to consult a detailed treatment of the inverse problem in geophysics (e.g., Parker 1994; Scales and others 2001; Menke 2012; Aster and others 2013). For the rest of us, Scales and others (2001) provide an engaging, nontechnical overview of the subject in the first three chapters of their book. Here we will briefly describe one example of inference.

Given a set of observational data, our objective is to find the best model (set of model parameters) m such that $d = G(m)$, where d and m are the data and model parameter vectors, respectively, and G is an operator that describes their relationship to one another. If this relationship is linear, the problem can be written as $d = Gm$, where G is called the observation matrix and is often constructed from Greens functions for that source. In the more general case, the relationship between the model and data is nonlinear, and G is a nonlinear operator; such problems can in some cases be linearized or else solved using any number of nonlinear optimization or statistical sampling techniques. For linear problems the best-fit model parameters are ideally obtained by $m = G^{-1}d$, so we need to invert G . Unfortunately, in volcano geodesy there is seldom enough data about the real Earth and its deformation to determine G well enough to simply invert, and consequently there are multiple possible solutions. To address this issue, Cervelli and others (2001) applied two methods for stochastic optimization, simulated annealing (Metropolis and others 1953) and random cost (Berg 1993), to the inversion of deformation data from the March 1997 earthquake swarm off the Izu Peninsula, Japan. They inverted GPS, tilt, and leveling data to arrive at a best-fit model involving dike intrusion and associated shear on a nearby fault.

Other approaches to solving the inverse problem involving deformation data include the network inversion filter introduced by Segall and Matthews (1997) and McGuire and Segall (2003). The method uses a stochastic description of local benchmark motion and an extended Kalman filter to smooth out the effects of noise

in deformation time series data. Kositsky and Avouac (2010) used a principal component analysis-based inversion method that can handle multiple deformation data sources over multiple timescales in a statistically rigorous way. To estimate spatially and temporally continuous ground deformation fields from repeated InSAR observations, Hetland and others (2012) developed multi-scale InSAR time series (MInTS) analysis, which combines wavelet decomposition in the spatial domain and a general parametrization in the time domain. The deformation field is estimated in the wavelet domain using a cross-validated, regularized least squares inversion.

Bayesian Inference

In the foregoing discussion, we placed no restriction on the range of parameter values to be considered, i.e., we invoked no prior knowledge of which values would be reasonable and which would not. In some cases, we might have independent information that can be used to place additional constraints on model parameter values. The problem with this is that both the model and the data are uncertain, and this uncertainty should be part of the solution. For example, if we are inverting gravity data to estimate the location and density of a subsurface magma body, we might have evidence that the magma's density is about 2000 kg/m^3 with a standard deviation of 1000 kg/m^3 . This knowledge can be formulated as a Gaussian probability density function and input to the model as prior knowledge. Furthermore, if past eruptions in the area produced only rhyolite and basalt flows in the ratio 1:10 and current gas emission data indicated a 70% likelihood that the source magma was basalt, we could use that information to guide the parameter search. In this simple example, the result presumably would be a high likelihood that the body is basaltic and a much lower (but non-zero) likelihood that it is rhyolitic. The use of such a priori knowledge is a key component of a statistical estimation method called Bayesian inference, named for the eighteenth century English statistician Thomas Bayes who derived the rule that bears his name (see Gelman and others 2013). Bayesian inference has been applied to various

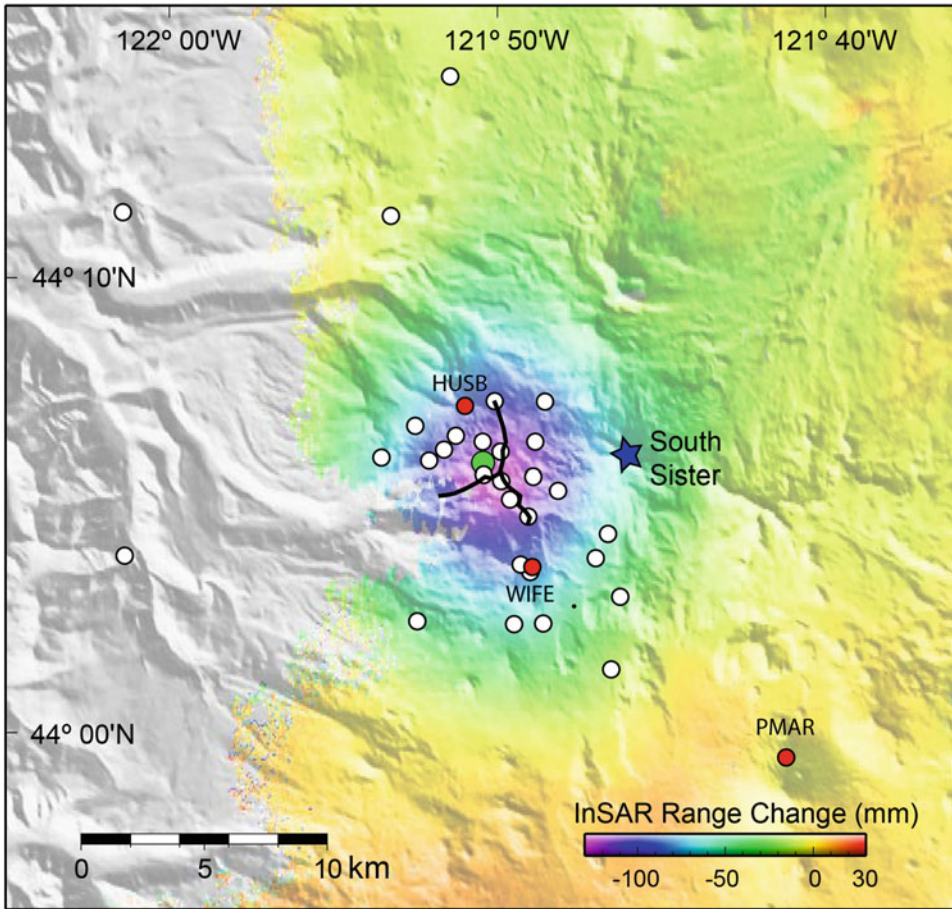
problems in volcanology, and its use is likely to increase as extensive multidisciplinary datasets become available for a larger number of volcanoes worldwide (e.g., Siebert and others 2010; Smithsonian Institution 2015). Trasatti and others (2009, 2011) used Bayesian source inference and 3D finite element models to study deformation episodes at Mount Etna (Italy) and the Campi Flegrei caldera (Italy), respectively. At Mount Etna, inversion of GPS, EDM, and InSAR data showed that inflation of the volcanic edifice from 1993 to 1997 was accompanied by sliding of the eastern flank, which was not detected by an earlier analysis. At Campi Flegrei, Bayesian inversion of leveling and EDM data together proved to be incompatible with an ellipsoid or crack source. Instead, Trasatti and others (2011) inferred a mixed mode (shear and tensile) dislocation at 5.5 km depth and reported that gravity changes measured during the event were compatible with the intrusion at that depth of $60\text{--}70 \times 10^6 \text{ m}^3$ of magma with density of $\sim 2400 \text{ kg/m}^3$ —details that might have escaped a less sophisticated analysis.

What Can Be Learned: Case Studies

Three Sisters Volcanic Center, Oregon

Three Sisters in the central Oregon Cascade Range is a long-lived center of basaltic to rhyolitic volcanism that includes three large composite cones of Quaternary age: Middle Sister, South Sister, and Broken Top. South Sister's most recent eruptions occurred in two closely spaced episodes between 2200 and 2000 years ago, producing rhyolite tephra, pyroclastic flows, lava flows, and lava domes (Scott 1987). As recently as 1600–1200 years ago, dominantly effusive eruptions of basaltic and andesitic lavas immediately north of Three Sisters built large shield volcanoes and isolated cinder cones and lava flows (Scott and others 2001; Sherrod and others 2004).

Repeated campaign-style GPS and leveling surveys, CGPS stations, and InSAR are being used to track an ongoing episode of surface uplift in the Three Sisters area that began in the late 1990s and was discovered with InSAR in 2001 (Fig. 6). In this case, wilderness restrictions made



Volcano Deformation: Insights into Magmatic Systems, Fig. 6 Map of CGPS (red, labeled) and campaign GPS (white) stations on a background of gray shaded relief and colored InSAR total range change from August 1995 to August 2001 at South Sister volcano, Oregon, USA (ERS interferogram from Dzurisin and others 2009). Green, blue, and violet hues in the interferogram represent negative range change, which indicates mostly surface

uplift in the case of a SAR with a steep look angle such as those aboard ERS-1 and ERS-2 (23° at mid-swath). Thick black lines near center mark level lines, and green circle near their intersection marks the center of uplift as determined from the GPS and level surveys. Blue star ~6 km east of the maximum uplift marks the summit of South Sister volcano

the installation of more than a few CGPS stations impractical. Instead, an existing campaign-style GPS network at South Sister volcano was expanded in 2001 to cover the entire deforming area, which is centered ~6 km west of the volcano's summit and spans ~10 km E-W $\times$ ~20 km N-S. Wicks and others (2002) used a Mogi source with corrections for topography (Williams and Wadge 1998) and tropospheric water vapor variation (Avallone and Briole 2000) to model three interferograms that collectively span 1995–2000,

and they concluded that surface deformation was caused by an inflation source at 6.5 ± 0.4 km depth (4.5 km below sea level, bsl). Dzurisin and others (2006) modeled InSAR, leveling, campaign-style GPS, and continuous GPS data that collectively span 1992–2004 and attributed the deformation to a dipping sill source at 6.5 ± 2.5 km depth (4.5 km bsl), with an average volume growth rate of 5×10^6 m³/year, starting sometime during 1996–1998. Based on observations through 2008, Dzurisin and others (2009) concluded that

the best-fit source model was a vertical, prolate spheroid at 4.9–5.4 km depth (2.9–3.4 km bsl) that inflated at an exponentially decreasing rate with a $1/e$ decay time of 5.3 ± 1.1 years starting in September 1997. Riddick and Schmidt (2011) modeled a more extensive InSAR dataset from 1992 to 2010 and inferred a more complex deformation history: Inflation of a source at ~5–7 km depth (3–5 km bsl) started gradually in mid-1996, increased significantly during 1998–2003, and then decreased in rate from 2004 to 2010. The decrease in inflation rate coincided with a swarm of small earthquakes under the deforming area in March 2004. Lisowski and others (2014) modeled all forms of deformation data available through the end of 2014 for the Three Sisters region (campaign GPS, CGPS, leveling, InSAR) and included in their model not only a volcanic source but also rigid block rotation and strain (McCaffrey 2009). They concluded that (1) the inflation rate of the volcanic source has been decaying exponentially since the event began sometime in 1998 and (2) a 7-km-deep (5 km bsl) spherical source provides the optimal fit to the data, given the level of uncertainty imposed by the data. The total source volume increase during the uplift episode, which had slowed considerably by the time this was written in January 2019, was $50\text{--}70 \times 10^6 \text{ m}^3$ (Riddick and Schmidt 2011; Lisowski and others 2014).

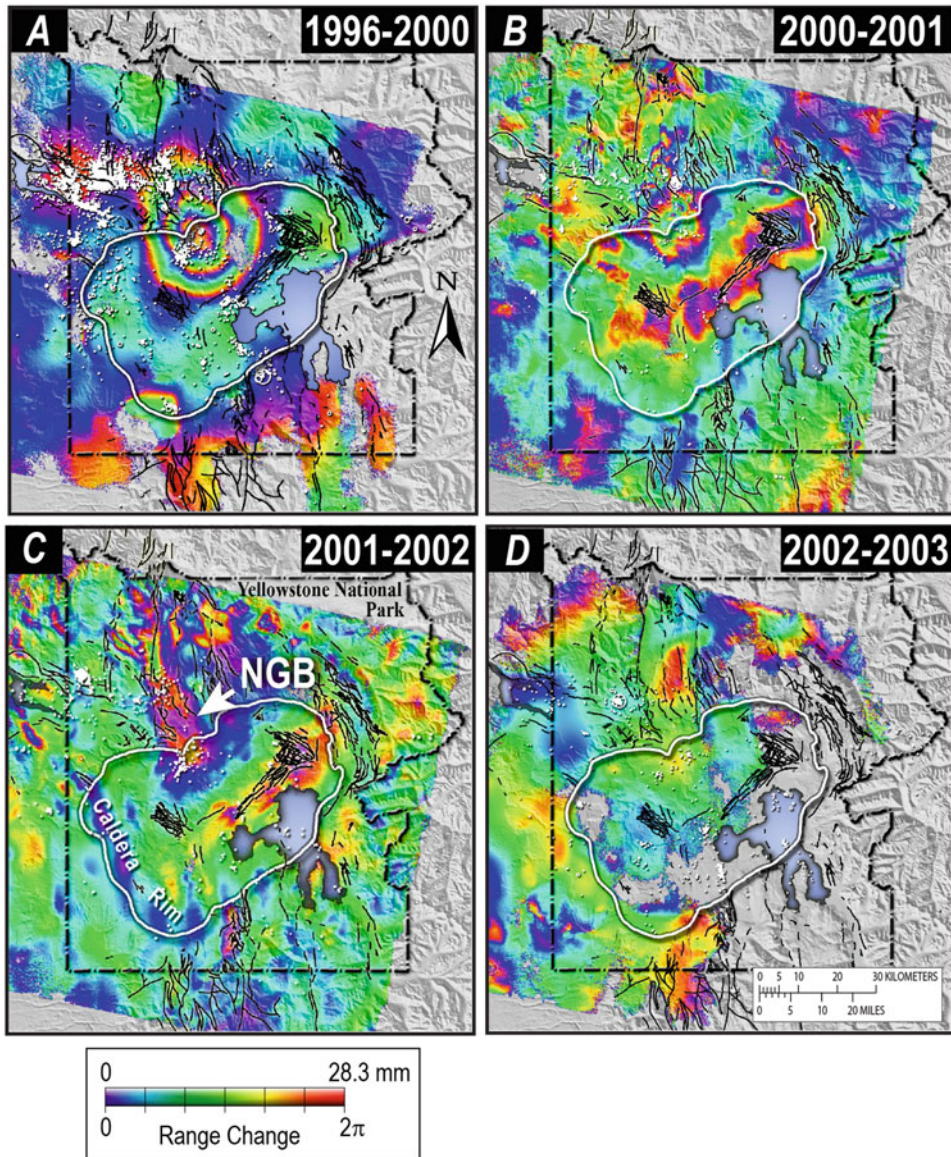
These results for Three Sisters illustrate both the complexity of real-world problems in volcano geodesy and the growing sophistication of analytical methods that can be brought to bear on them. As time passes, both more data and more types of data become available to constrain the deformation source and its history. Important details remain to be resolved, but the location, depth, and decaying time history of the Three Sisters inflation source are well established as this is written (January 2019).

Yellowstone Magmatic System, Wyoming

The Yellowstone magmatic system has produced three caldera-forming eruptions in the past 2.1 million years, and the youngest of the calderas, 640,000-year-old Yellowstone Caldera, currently

is among the most restless caldera systems on Earth (Christiansen 2001). Yellowstone is remarkable for its high levels of swarm seismicity (Farrell and others 2009), hydrothermal activity (Hurwitz and Lowenstern 2014), magmatic gas emission (Werner and Brantley 2003), and ground deformation (Dzurisin and others 2012). Several modeling approaches have been taken, and various source mechanisms have been proposed to explain the observed ground deformation, which varies in both space and time (e.g., Wicks and others 2006; Chang and others 2007; Vasco and others 2007; Dzurisin and others 2012) (Fig. 7). Common elements in all the models are a flux of basaltic magma from the mantle, a partly molten crustal reservoir of silicic magma remnant from past eruptions, and an overlying hydrothermal system that can self-seal and pressurize to cause surface uplift or rupture to cause subsidence. Gases released from magma as it ascends or cools and brines that form in the deep hydrothermal system can play a role in surface deformation if they become trapped beneath the self-sealed layer (Fournier 2007). Over long timescales, poroelastic and thermoelastic deformation of rock hosting the hydrothermal system may also be important (Hutnak and others 2009). A key unresolved issue at Yellowstone is the role played by and relative importance of magma, hydrous fluid, and magmatic gases in causing ground deformation.

The most direct way to address this issue is to measure the density ρ of the fluid involved in deformation. If $\rho \geq 2000 \text{ kg/m}^3$, the fluid is almost surely magma. If $1000 \text{ kg/m}^3 \leq \rho \leq 1500 \text{ kg/m}^3$, water or brine is the most likely deformation agent. Finally, if $\rho < 1000 \text{ kg/m}^3$, the agent is either magmatic gas or steam, or the deformation is thermoelastic with little or no fluid involved. Smith and others (1989) noted that the relationship between height change and gravity change, $\Delta g/\Delta h$, determined by repeated leveling and microgravity surveys from 1977 to 1987 was significantly less than the normal free-air gradient (-0.117 mGal/m versus -0.3086 mGal/m , respectively). The observed relationship suggests that a net mass increase occurred beneath the caldera during a



Volcano Deformation: Insights into Magmatic Systems, Fig. 7 Four ERS-2 interferograms superimposed on digital terrain showing range changes at the Yellowstone Caldera, Yellowstone National Park, between 1996 and 2003. A color change from violet to blue to green to yellow to red, shown in the color bar, marks an increase in the range (mostly subsidence in this case) of 28.3 mm; the opposite color progression indicates uplift. White dots represent epicenters of earthquakes recorded during the time interval spanned by each interferogram. (a) Summer 1996 to summer 2000 interferogram. White outline shows caldera rim. Concentric pattern of fringes indicates uplift along the north caldera rim. Although the caldera floor appears to have subsided only slightly, this period includes about 30 mm of caldera-wide uplift from 1995 to 1997

(Wicks and others 1998). Therefore, more than 30 mm of subsidence of the caldera floor occurred in the central part of the caldera from 1997 to 2000, while the north rim rose about 80 mm. (b) Summer 2000 to summer 2001 interferogram showing continuing uplift along the north caldera rim and subsidence of the caldera floor. (c) Summer 2001 to summer 2002 interferogram. Uplift along the north rim continued and extended northward, while the caldera floor continued to subside at a declining rate. Arrowlabeled NGB marks the location of Norris Geyser Basin. (d) Summer 2002 to summer 2003 interferogram showing slight subsidence of the north rim area and northward, and no measurable motion of the caldera floor. (Figure from Wicks and others 2006)

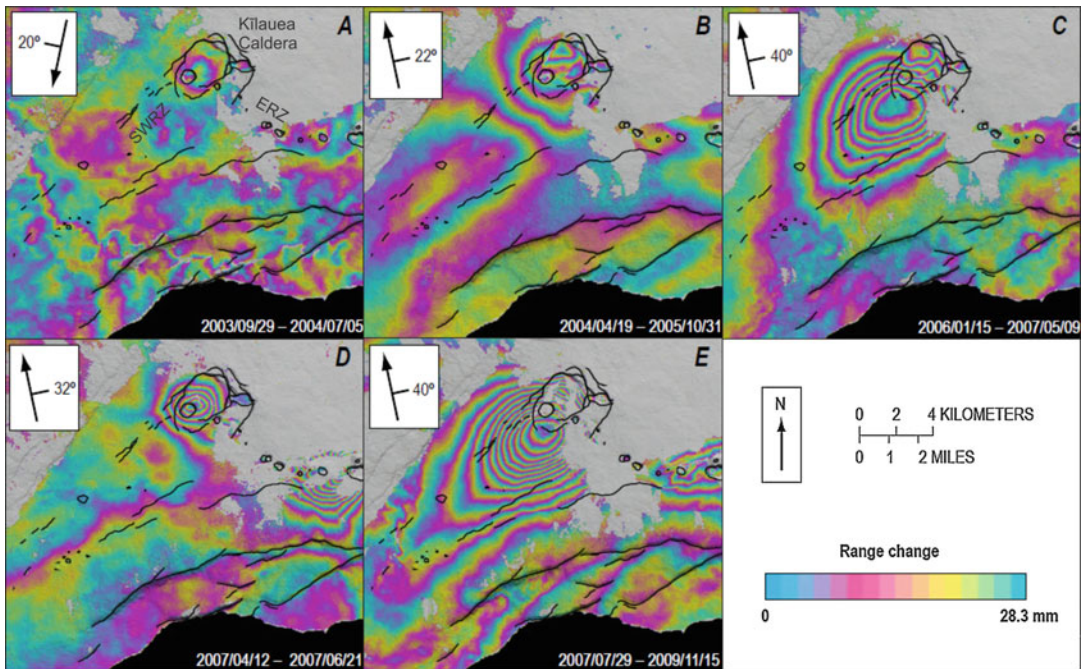
period of net caldera floor uplift (Dzurisin and others 2012). Arnet and others (1997) extended the gravity observations through 1994 to include more of a subsidence episode that began in 1985. They discovered that the gravity gradient during subsidence was very close to the free-air gradient, implying little or no mass change. Arnet and others (1997, p. 2744) attributed the earlier mass increase to “...widespread hydrothermal fluid movement, which furthermore is related to input by magma” and the subsequent lack of mass change during subsidence to “...depressurization of the deep hydrothermal system as a result of fracturing and volatile loss to the shallow hydrothermal system...” This explanation is appealing because it accounts for alternating periods of uplift and subsidence, which have been observed several times at Yellowstone, as a consequence of a continuum process involving basalt intrusion and fluid accumulation (mass increase) followed by degassing and volatile loss (no mass change). On the other hand, the result has yet to be confirmed since the discovery of another locus of deformation centered along the north caldera rim near Norris Geyer Basin (Wicks and others 2006; Puskas and others 2007). The north rim area generally subsides, while the caldera floor rises, and vice versa. Such was the case during a period of rapid caldera uplift during 2004–2008 (Chang and others 2007), and during an episode of north rim subsidence and caldera uplift that began soon after a magnitude 4.8 earthquake in the Norris area on March 30, 2014 (Stovall and others 2014). It has been suggested that the caldera and north rim sources are hydraulically connected, and their generally inverse behavior is indicative of fluid exchange between them (e.g., Wicks and others 2006; Puskas and others 2007). Additional gravity observations are needed to further test this intriguing idea.

Kīlauea and Mauna Loa Volcanoes, Hawai‘i

Deformation studies at Kīlauea and Mauna Loa volcanoes have been ongoing for more than a century, and much has been learned about the dynamics of Hawaiian volcanism as a result. Geodesy has been a key to the discovery of shallow magma reservoirs beneath the summits of both volcanoes and along active rift zones, to

recognition of rapid seaward motion of Kīlauea’s south flank, and to estimates of the magma supply rate (see Decker 1987, and references therein). In a recent synthesis, Poland and others (2014) described what is known about magma supply, storage, and transport at Hawaiian volcanoes. They combined deformation, seismic, gas emission, petrologic, and effusion rate data to argue for a mantle-driven pulse in magma supply to Kīlauea during 2003–2007 (see also Poland and others 2012) (Fig. 8). Among the consequences of the increased supply rate, Poland and others (2014, p. 189) listed (1) rapid summit inflation and an intrusion of magma from beneath the summit into the East Rift Zone (ERZ) that (a) resulted in a small eruption there in June 2007, (b) increased compressive stress on Kīlauea’s south flank, and (c) triggered an aseismic slip event on a basal detachment fault shortly after the start of the intrusion; (2) formation of a new long-term eruptive vent about 2 km downrift of the previously active Pu‘u ‘Ō‘ō vent, which had been the dominant source of lava effusion since 1992 (and also during 1983–1986); and (3) formation of a vent and persistent lava lake in the summit area in March 2008. They also noted that during approximately the same time period, (4) nearby Mauna Loa volcano began to inflate in 2002 after nearly a decade of deflation; (5) the inflation rate increased in late 2004, accompanied by thousands of long-period earthquakes at >30-km depth; and (6) inflation waned and ceased by the end of 2009 (Fig. 9). Simultaneous inflation of both volcanoes accompanied by a swarm of deep long-period earthquakes suggests a temporary increase in magma supply from a source region in the mantle. Deformation studies alone cannot “see” so deeply into the Earth, but together with information from other disciplines, they argue for a deep connection between Kīlauea and Mauna Loa, the existence and nature of which have been debated for more than a century (see discussion on pp. 189–190 of Poland and others 2014, and references therein).

A multidisciplinary approach involving geodesy, seismology, geology, geochemistry, and geophysics likewise has advanced our understanding of the magma plumbing system at Kīlauea. On



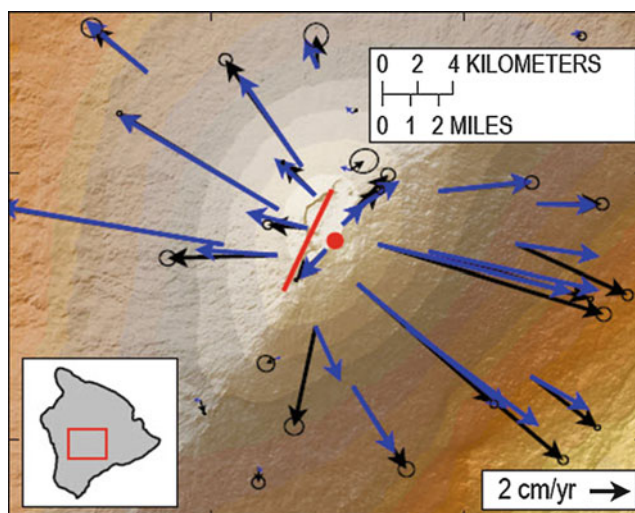
Volcano Deformation: Insights into Magmatic Systems, Fig. 8 Interferograms showing surface deformation of Kilauea's summit area associated with a magma supply increase during 2003–2007. Data are from the Advanced Synthetic Aperture Radar (ASAR) instrument on the Envisat satellite. For all images, dates spanned are in the lower right, and upper left inset gives flight direction (arrow) and look direction (orthogonal line with angle in degrees from vertical). Color scale in the lower right panel applies to all five interferograms. One fringe is 28.3 mm of range change along the radar line of sight, where positive (magenta to yellow) indicates increasing range, that is, ground motion away from the satellite (mostly

subsidence). (a) Uplift is focused within Kilauea Caldera, near Halema'uma'u Crater (circular pit crater in south part of caldera). (b) Uplift has shifted a few kilometers south-eastward to near Keanakāko'i Crater along the upper East Rift Zone (ERZ). (c) Uplift rate has increased and deformation is centered on the south caldera rim and upper Southwest Rift Zone (SWRZ). (d) Subsidence near Halema'uma'u Crater and uplift of the ERZ was caused by magma withdrawal from the summit and intrusion into the ERZ during June 17–19, 2007. (e) Subsidence centered in the south caldera and upper SWRZ, as well as in the ERZ. (Figure from Poland and others 2014)

that basis, Poland and others (2014, pp. 191–210) proposed a refined model consisting of (1) a primary storage reservoir ~3–5 km beneath the south caldera, (2) a smaller reservoir ~1–2 km beneath the east margin of Halema'uma'u Crater (a nested pit crater within the caldera), (3) an intermittent storage zone near the intersection of the summit caldera rim and ERZ, (4) pathways and subsidiary storage zones along the ERZ and Southwest Rift Zone (SWRZ) that radiate from the primary summit reservoir, (5) shallower pathways that extend from the smaller summit reservoir to the upper parts of both rift zones, and (6) a poorly understood “deep rift zone” at ~3–9 km depth beneath the ERZ and SWRZ.

Future Directions

The case studies discussed above illustrate both the broad range of topics that can be addressed by geodetic investigations at volcanoes and the utility of combining geodetic information with that from other disciplines. In the foreseeable future, data from high-quality multidisciplinary sensor networks will become increasingly available in real or near-real time. Analysis and fusion of the disparate datasets will be done routinely, and physics-based models will be updated with the results in near-real time. Techniques for rapid processing of CGPS data with latencies approaching 1 s have been developed and are



Volcano Deformation: Insights into Magmatic Systems, Fig. 9 Surface displacements near the summit of Mauna Loa volcano during 2004–2005 from GPS data. Observed displacements are black, with 95% confidence ellipses; modeled displacements are blue. Model includes a point (Mogi) source at 3.5 km depth beneath the southeast

caldera rim (red dot), with a volume increase of $4 \times 10^6 \text{ m}^3/\text{year}$, and a vertical dike along the length of the caldera (red line), with a top at 3.8 km depth and a volume increase of $30 \times 10^6 \text{ m}^3/\text{year}$. (Figure from Poland and others 2014)

being implemented for both commercial and research purposes. In the second category, Cannavò and others (2015) described a method for real-time tracking of magmatic intrusions by inversion of high-rate CGPS data and demonstrated the method's utility by simulated application to an intrusive event at Mount Etna. D'Auria and others (2012) used a tomographic imaging technique to invert InSAR data for the Campi Flegrei caldera and to identify three episodes of fluid injection along the caldera margin during 1995–2007. Their 4D analysis showed that fluid migrated upward and toward the center of the caldera during two of the episodes – the type of information that will be available to support timely assessments of evolving hazards at Campi Flegrei and elsewhere in the near future.

In addition to more timely analysis and fusion of volcano monitoring datasets, physics-based models that can ingest diverse datasets and evolve accordingly over time are another promising research topic in volcanology. Anderson and Segall (2011) applied such a model to the 2004–2008 dome-building eruption of Mount St. Helens, Washington, and concluded that little

or no recharge of the magma reservoir feeding the eruption occurred while the eruption was in progress – information that, if available in near-real time, could in the future support updated hazard assessments and forecasts of when an eruption is likely to end or intensify. For the same eruption, Anderson and Segall (2013) used a physics-based model and Bayesian inversion technique to estimate several characteristics of the Mount St. Helens magma reservoir that are difficult to constrain by more conventional means, including reservoir volume, overpressure, and volatile content. Detailed information about surface deformation, seismicity, gas emission, and microgravity is essential to improving models in the future.

The coming decade will see additional launches of advanced GPS and SAR satellites, improved techniques for real-time analysis, fusion, and modeling of disparate volcano datasets, and more widespread and effective communication of volcano hazards. Volcano geodesy will play an important role in each of these areas, to the betterment of volcano science and the ever-increasing number of people at risk from future volcanic activity.

Acknowledgments The author was supported by the U.S. Geological Survey's Volcano Hazards Program through its Volcano Science Center and David A. Johnston Cascades Volcano Observatory.

Bibliography

Primary Literature

- Agram PS, Jolivet R, Riel B, Lin YN, Simons M, Hetland E, Doin MP, Lasserre C (2013) New radar interferometric time series analysis toolbox released. *EOS Trans Am Geophys Union* 94:69–70. <https://doi.org/10.1002/2013EO070001>
- Agustsson K, Stefansson R, Linde AT, Einarsson P, Sacks IS, Gudmundsson GB, Thorbjarnsdottir B (2000) Successful prediction and warning of the 2000 eruption of Hekla based on seismicity and strain changes. *Eos Trans Am Geophys Union* 81(48). Fall Meeting Supplement, Abstract F1337
- Ali ST, Feigl KL, Carr BB, Masterlark T, Sigmundsson F (2014) Geodetic measurements and numerical models of rifting in Northern Iceland for 1993–2008. *Geophys J Int* 196:1267–1280. <https://doi.org/10.1093/gji/ggt462>
- Anderson EM (1936) Dynamics of the formation of cone-sheets, ring-dykes, and cauldron-subsidences. *Proc R Soc Edinburgh* 56:128–157
- Anderson K, Segall P (2011) Physics-based models of ground deformation and extrusion rate at effusively erupting volcanoes. *J Geophys Res* 116, :B07204, 20. <https://doi.org/10.1029/2010JB007939>
- Anderson K, Segall P (2013) Bayesian inversion of data from effusive volcanic eruptions using physics-based models; application to Mount St. Helens 2004–2008. *J Geophys Res* 118(5):2017–2037. <https://doi.org/10.1002/jgrb.50169>
- Arnet F, Kahle H-G, Klingelé E, Smith RB, Meertens CM, Dzurisin D (1997) Temporal gravity and height changes of Yellowstone caldera, 1977–1994. *Geophys Res Lett* 24(22):2741–2744
- Avallone A, Briole P (2000) Analysis of 5 years of SAR interferometry data from the Gulf of Corinth (Greece). *EOS Trans Am Geophys Union* 81:F327
- Bagnardi M, Poland MP, Carbone D, Baker S, Battaglia M, Amelung F (2014) Gravity changes and deformation at Kīlauea Volcano, Hawai‘i, associated with summit eruptive activity, 2009–2012. *J Geophys Res* 119(9): 7288–7305. <https://doi.org/10.1002/2014JB011506>
- Battaglia M, Segall P, Roberts C (2003) The mechanics of unrest at Long Valley caldera, California; 2. Constraining the nature of the source using geodetic and micro-gravity data. In: Sorey ML, McConnell VS, Roeloffs E (eds) *Crustal Unrest in Long Valley Caldera, California: new interpretations from geophysical and hydrologic monitoring and deep drilling*. Elsevier, Amsterdam. *J Volcanol Geotherm Res* 127(3–4): 219–245
- Berg B (1993) Locating global minima in optimization problems by a random-cost approach. *Nature* 361: 708–710
- Berrino G, Corrado G, Riccardi U (2006) On the capability of recording gravity stations to detect signals coming from volcanic activity: the case of Vesuvius. *J Volcanol Geotherm Res* 150:270–282. <https://doi.org/10.1016/j.jvolgeores.2005.07.015>
- Biggs J, Wright T, Lu Z, Parsons B (2007) Multi-interferogram method for measuring interseismic deformation: Denali Fault, Alaska. *Geophys J Int* 170:1165–1179. <https://doi.org/10.1111/j.1365-246X.2007.03415.x>
- Biggs J, Lu Z, Fournier T, Freymueller JT (2010) Magma flux at Okmok Volcano, Alaska, from a joint inversion of continuous GPS, campaign GPS, and interferometric synthetic aperture radar. *J Geophys Res* 115(B12401):11. <https://doi.org/10.1029/2010JB007577>
- Blewitt G, Hammond WC, Kreemer C (2009) Geodetic observation of contemporary deformation in the northern Walker Lane: 1. Semipermanent GPS strategy. In: Oldow JS, Cashman PH (eds) *Late Cenozoic structure and evolution of the Great Basin–Sierra Nevada transition*, Geological Society of America special paper, vol 447. Geological Society of America, Boulder, pp 1–15. [https://doi.org/10.1130/2009.2447\(01\)](https://doi.org/10.1130/2009.2447(01))
- Bonaccorso A, Davis PM (1999) Models of ground deformation from vertical volcanic conduits with application to eruptions of Mount St. Helens and Mount Etna. *J Geophys Res* 104(B5):10531–10542
- Bonaccorso A, Currenti G, Del Negro C (2013) Interaction of volcano-tectonic fault with magma storage, intrusion and flank instability; a thirty years study at Mt. Etna volcano. *J Volcanol Geotherm Res* 251:127–136
- Bonafede M, Ferrari C (2009) Analytical models of deformation and residual gravity changes due to a Mogi source in a viscoelastic medium. *Tectonophysics* 471:4–13
- Bonforte A, Colesanti C, Ferretti A, Guglielmino F, Palano M, Prati C, Puglisi G, Rocca F (2004) Remote sensing for ground deformation analysis during the eruptive event of July 2001 at Mt. Etna, paper presented at ESA SP-550. FRINGE 2003 Workshop, 1 June 2004
- Brown GC, Rymer H (1991) Microgravity monitoring of active volcanoes: a review of theory and practice. *Cahiers du Centre European de Geodynamique et de Seismologie* 4:279–304
- Canitano A, Bernard P, Linde AT, Sacks S, Boudin F (2014) Correcting high-resolution borehole strainmeter data from complex external influences and partial-solid coupling; the case of Trizonia, Rift of Corinth (Greece). *Pure Appl Geophys* 171(8):1759–1790
- Cannavò F, Camacho AG, González PJ, Mattia M, Puglisi G, Fernández J (2015) Real time tracking of magmatic intrusions by means of ground deformation modeling during volcanic crises. *Scientific Reports* 5:10970. <https://doi.org/10.1038/srep10970>
- Caputo M (1979) Two-thousand years of geodetic and geophysical observations in the Phlegraean fields near Naples. *Geophys J R Astron Soc* 56:319–328

- Carbone D, Poland MP (2012) Gravity fluctuations induced by magma convection at Kīlauea volcano, Hawai'i. *Geology* 40:803–806
- Carbone D, Budetta G, Greco F (2003a) Possible mechanisms of magma redistribution under Mt. Etna during the 1994–1999 period detected through microgravity measurements. *Geophys J Int* 153:187–200
- Carbone D, Budetta G, Greco F (2003b) Bulk processes prior to the 2001 Mount Etna eruption, highlighted through microgravity studies. *J Geophys Res* 108(B12). <https://doi.org/10.1029/2003JB002542>
- Carbone D, Poland MP, Patrick MR, Orr TR (2013) Continuous gravity measurements reveal a low-density lava lake at Kīlauea Volcano, Hawai'i. *Earth Planet Sci Lett* 376:178–185. <https://doi.org/10.1016/j.epsl.2013.06.024>
- Cervelli P, Murray MH, Segall P, Aoki Y, Kato T (2001) Estimating source parameters from deformation data, with an application to the March 1997 earthquake swarm off the Izu Peninsula, Japan. *J Geophys Res* 106(B6):11217–11237. <https://doi.org/10.1029/2000JB900399>
- Chang W-L, Smith RB, Wicks C, Farrell JM, Puskas CM (2007) Accelerated uplift and magmatic intrusion of the Yellowstone caldera, 2004 to 2006. *Science* 318(5852):952–956. <https://doi.org/10.1126/science.1146842>
- Chouet B (2003) Volcano seismology. *Pure Appl Geophys* 160:739–788. <https://doi.org/10.1007/PL00012556>
- Christiansen RL (2001) The Quaternary and Pliocene Yellowstone Plateau volcanic field of Wyoming, Idaho, and Montana. U.S. Geological survey professional paper 729-G, p 145, 3 plates, scale 1:125,000
- Collins BD, Kayen RE (2006) Land-based lidar mapping – a new surveying technique to shed light on rapid topographic change. US Geological Survey Fact Sheet 2006–3111, p 4. <http://pubs.usgs.gov/fs/2006/3111/>
- Currenti G, Bonaccorso A, Del Negro C, Scandura D, Boschi E (2010) Elasto-plastic modeling of volcano ground deformation. *Earth Planet Sci Lett* 296(3–4): 311–318
- D'Auria L, Giudicepietro F, Martini M, Lanari R (2012) The 4D imaging of the source of ground deformation at Campi Flegrei caldera (southern Italy). *J Geophys Res* 117:B08209. <https://doi.org/10.1029/2012JB009181>
- Davis PM (1983) Surface deformation associated with a dipping hydrofracture. *J Geophys Res* 88(B7): 5826–5834
- Davis PM (1986) Surface deformation due to inflation of an arbitrarily oriented triaxial ellipsoidal cavity in an elastic half-space, with reference to Kīlauea Volcano, Hawai'i. *J Geophys Res* 91:7429–7438
- de Zeeuw-van Dalsen E, Rymmer H, Sigmundsson F, Sturkell E (2005) Net gravity decrease at Askja Volcano, Iceland; constraints on processes responsible for continuous caldera deflation, 1988–2003. *J Volcanol Geotherm Res* 139:227–239
- Decker RW (1987) Dynamics of Hawaiian volcanoes: an overview, chapter 42. In: Decker RW, Wright TL, Stauffer PH (eds) *Volcanism in Hawaii*, pp 997–1018. US Geological Survey Professional Paper 1350
- Del Negro C, Currenti G, Scandura D (2009) Temperature-dependent viscoelastic modeling of ground deformation; application to Etna volcano during the 1993–1997 inflation period. *Phys Earth Planet Inter* 172:299–309
- Diefenbach AK, Crider JG, Schilling SP, Dzurisin D (2012) Rapid, low-cost photogrammetry to monitor volcanic eruptions: an example from Mount St. Helens, Washington, USA. *Bull Volcanol* 74(2):579–587. <https://doi.org/10.1007/s00445-011-0548-y>. <http://link.springer.com/article/10.1007%2Fs00445-011-0548-y>
- Dieterich JH, Decker RW (1975) Finite element modeling of surface deformation associated with volcanism. *J Geophys Res* 80:4095–4102
- Dzurisin D, Anderson LA, Eaton GP, Koyanagi RY, Okamura RT, Puniwai GS, Sako MK, Yamashita KM (1980) Geophysical observations of Kīlauea volcano, Hawai'i; 2. Constraints on the magma supply during November 1975–September 1977. *J Volcanol Geotherm Res* 7:241–270
- Dzurisin D, Lisowski M, Wicks CW, Poland MP, Endo ET (2006) Geodetic observations and modeling of magmatic inflation at the three sisters volcanic center, Central Oregon Cascade Range, USA. *J Volcanol Geotherm Res* 150(1–3):35–54. <https://doi.org/10.1016/j.jvolgeores.2005.07.011>. Special issue The changing shape of active volcanoes
- Dzurisin D, Lisowski M, Wicks CW (2009) Continuing inflation at three sisters volcanic center, Central Oregon Cascade Range, USA, from GPS, leveling, and InSAR observations. *Bull Volcanol* 71:1091–1110. <https://doi.org/10.1007/s00445-009-0296-4>
- Dzurisin D, Wicks CW, Poland MP (2012) History of surface displacements at the Yellowstone Caldera, Wyoming, from leveling surveys and InSAR observations, 1923–2008. U.S. Geological survey professional paper 1788, p 68 and data files. Available at <http://pubs.usgs.gov/pp/1788/>
- Dzurisin D, Lisowski M, Wicks CW Jr (2017) Semipermanent GPS (SPGPS) as a volcano monitoring tool: rationale, method, and applications. *J Volcanol Geotherm Res* 344:40–51. <https://doi.org/10.1016/j.jvolgeores.2017.03.007>. <https://doi.org/10.1016/j.jvolgeores.2017.03.007>
- Farrell J, Husen S, Smith RB (2009) Earthquake swarm and *b*-value characterization of the Yellowstone volcano-tectonic system. *J Volcanol Geotherm Res* 188:260–276
- Favalli M, Fornaciai A, Mazzarini F, Harris A, Neri M, Behncke B, Pareschi MT, Tarquini S, Boschi E (2010) Evolution of an active lava flow field using a multi-temporal LIDAR acquisition. *J Geophys Res* 115. <https://doi.org/10.1029/2010JB007463>
- Feigl KL, Thurber CH (2009) A method for modelling radar interferograms without phase unwrapping: application to the M 5 Fawnskin, California earthquake of 1992 December 4. *Geophys J Int* 176(2):491–504
- Ferretti A, Prati C, Rocca F (2000) Nonlinear subsidence rate estimation using permanent scatterers in differential SAR interferometry. *IEEE Trans Geosci Remote Sens* 38:2202–2212

- Ferretti A, Prati C, Rocca F (2001) Permanent scatterers in SAR interferometry. *IEEE Trans Geosci Remote Sens* 39:8–20
- Ferretti A, Fumagalli A, Novali F, Prati C, Rocca F, Rucci A (2011) A new algorithm for processing interferometric data-stacks: SqueeSAR. *IEEE Trans Geosci Remote Sens* 49(9):3460–3470. <https://doi.org/10.1109/TGRS.2011.2124465>
- Fialko Y, Khazan Y, Simons M (2001) Deformation due to a pressurized horizontal circular crack in an elastic half-space, with applications to volcano geodesy. *Geophys J Int* 146:181–190
- Foulger GR, Julian BR, Hill DP, Pitt AM, Malin PE, Shalev E (2004) Non-double-couple microearthquakes at Long Valley caldera, California, provide evidence for hydraulic fracturing. *J Volcanol Geotherm Res* 132:45–71
- Fournier RO (2007) Hydrothermal systems and volcano geochemistry, chapter 10. In: Dzurisin D (ed) *Volcano deformation – geodetic monitoring techniques*, Springer-Praxis books in geophysical sciences. Springer, Berlin, p 441
- Furuya M, Okubo S, Sun W, Tanaka Y, Oikawa J, Watanabe H, Maekawa T (2003) Spatiotemporal gravity changes at Miyakejima volcano, Japan; Caldera collapse, explosive eruptions and magma movement. *J Geophys Res* 108(B4):2219. <https://doi.org/10.1029/2002JB0011989>
- Gerlach TM, McGee KA, Elias T, Sutton AJ, Doukas MP (2002) Carbon dioxide emission rate of Kilauea Volcano; Implications for primary magma and the summit reservoir. *J Geophys Res* 107:2189. <https://doi.org/10.1029/2001JB000407>
- Hammond WC, Kreemer C, Blewitt G (2007) Exploring the relationship between geothermal resources and geodetically inferred faults slip rates in the Great Basin. *Geotherm Resour Counc Trans* 31:391–395
- Hetland EA, Musé P, Simons M, Lin YN, Agram PS, DiCaprio CJ (2012) Multiscale InSAR time series (MInTS) analysis of surface deformation. *J Geophys Res* 117:B02404. <https://doi.org/10.1029/2011JB008731>
- Hickey J, Gottsmann JH (2014) Benchmarking and developing numerical finite element models of volcanic deformation. *J Volcanol Geotherm Res* 280:126–130. <https://doi.org/10.1016/j.jvolgeores.2014.05.011>
- Hole JK, Hooper A, Wadge G, Stevens NF, Anonymous (2006) Measuring contemporary deformation in the Taupo volcanic zone, New Zealand, using SAR interferometry. In: ESA SP, vol 610, p 6. European Space Agency (ESA) Scientific & Technical Publications, Noordwijk
- Hooper A (2008) A multi-temporal InSAR method incorporating both persistent scatterer and small baseline approaches. *Geophys Res Lett* 35. <https://doi.org/10.1029/2008GL034654>
- Hooper A, Zebker HA (2007) Phase unwrapping in three dimensions with application to InSAR time series. *J Opt Soc Am A* 24:2737–2747
- Hooper A, Zebker H, Segall P, Kampes B (2004) A new method for measuring volcanic deformation using InSAR persistent scatterers. *Geophys Res Lett* 31: L23611. <https://doi.org/10.1029/2004GL021737>
- Hooper A, Segall P, Zebker H (2007) Persistent scatterer interferometric synthetic aperture radar for crustal deformation analysis, with application to Volcán Alcedo Galápagos. *J Geophys Res* 112:21. <https://doi.org/10.1029/2006JB004763>
- Hunt T, Sugihara M, Sato T, Takemura T (2002) Measurement and use of the vertical gravity gradient in correcting repeat microgravity measurements for the effects of ground subsidence in geothermal systems. *Geothermics* 31:525–543
- Hutnak M, Hurwitz S, Ingebritsen SE, Hsieh PA (2009) Numerical models of caldera deformation; effects of multiphase and multicomponent hydrothermal fluid flow. *J Geophys Res* 114:B04411. <https://doi.org/10.1029/2008JB006151>
- Jachens RC, Eaton GP (1980) Geophysical observations of Kilauea volcano, Hawai'i; 1. Temporal gravity variations related to the 29 November, 1975, M = 7.2 earthquake and associated summit collapse. *J Volcanol Geotherm Res* 7:225–240
- Johnson DJ (1995) Gravity changes on Mauna Loa volcano. In: Rhodes JM, Lockwood JP (eds) *Mauna Loa revealed: structure, composition, history and hazards*, Geophysical monograph, vol 92. American Geophysical Union, Washington DC, pp 127–193
- Johnson DJ, Eggers AA, Bagnardi M, Battaglia M, Poland MP, Miklius A (2010) Shallow magma accumulation at Kilauea Volcano, Hawai'i, revealed by microgravity surveys. *Geology* 38(12):1139–1142. <https://doi.org/10.1130/G31323.1>
- Jousset P, Dwipa S, Beauducel F, Duquesnoy T, Diament M (2000) Temporal gravity at Merapi during the 1993–1995 crisis; an insight into the dynamical behavior of volcanoes. *J Volcanol Geotherm Res* 100(1–4):289–320
- Kauahikaua JP, Miklius A (2003) Long-term trends in microgravity at Kilauea's summit during the Pu'u 'O'o-Kupaianaha eruption. In: Heliker C, Swanson DA, Takahashi TJ (eds) *The Pu'u 'O'o-Kupaianaha eruption of Kilauea Volcano, Hawai'i: The first 20 years*. U.S. Geological survey professional paper, vol 1676, pp 165–171. http://pubs.usgs.gov/pp/pp1676/pp1676_10.pdf
- Kositsky AP, Avouac J-P (2010) Inverting geodetic time series with a principal component analysis-based inversion method. *J Geophys Res* 115:B03401. <https://doi.org/10.1029/2009JB006535>
- Linde AT, Sacks IS (2000) Real time predictions of imminent volcanic activity using borehole deformation data. *Eos Trans Am Geophys Union* 81(48). Fall Meeting Supplement, Abstract F1253
- Linde AT, Agustsson K, Sacks IS, Stefansson R (1993) Mechanism of the 1991 eruption of Hekla from continuous borehole strain monitoring. *Nature* 365:737–740

- Lipman PW, Lockwood JP, Okamura RG, Swanson DA, Yamashita KM (1985) Ground deformation associated with the 1975 magnitude-7.2 earthquake and resulting changes in activity of Kilauea Volcano, Hawai'i. *US Geol Surv Prof Pap* 1276:1–45
- Lisowski M (2007) Analytical deformation source models, chapter 8. In: Dzurisin D (ed) *Volcano deformation – geodetic monitoring techniques*, Springer-Praxis books in geophysical sciences. Springer, Berlin, pp 279–304
- Lisowski M, McCaffrey R, Wicks C, Dzurisin D (2014) Decaying rate of volcanic inflation near Three Sisters, Oregon, measured with GPS and InSAR [poster], 2014 Fall AGU Meeting, San Francisco, V41B-4795
- Massonnet D, Feigl KL (1995) Discriminating geophysical phenomena in satellite radar interferograms. *Geophys Res Lett* 22:1537–1540
- Massonnet D, Rossi M, Carmona C, Adragna F, Peltzer G, Feigl K, Rabaute T (1993) The displacement field of the Landers earthquake mapped by radar interferometry. *Nature* 364:138–142
- Massonnet D, Briole P, Arnaud A (1995) Deflation of Mount Etna monitored by spaceborne radar interferometry. *Nature* 375:567–570
- Masterlark T (2007) Magma intrusion and deformation predictions: sensitivities to the Mogi assumptions. *J Geophys Res* 112:B06419. <https://doi.org/10.1029/2006JB004860>
- Masterlark T, Haney M, Dickinson H, Fournier T, Searcy C (2010) Rheologic and structural controls on the deformation of Okmok volcano, Alaska; FEMs, InSAR, and ambient noise tomography, B02409. *J Geophys Res* 115. <https://doi.org/10.1029/2009JB006324>
- Masterlark T, Feigl KL, Haney M, Stone J, Thurber C, Ronchin E (2011) Nonlinear estimation of geometric parameters in FEMs of volcano deformation; integrating tomography models and geodetic data for Okmok volcano, Alaska. *J Geophys Res Atmos* 117(B2):2407. <https://doi.org/10.1029/2011JB008811>
- Masterlark T, Feigl KL, Haney M, Stone J, Thurber C, Ronchin E (2012) Nonlinear estimation of geometric parameters in FEMs of volcano deformation; integrating tomography models and geodetic data for Okmok volcano Alaska. *J Geophys Res Solid Earth* 117: B020407. <https://doi.org/10.1029/2011JB008811>
- McCaffrey R (2009) Time-dependent inversion of three-component continuous GPS for steady and transient sources in northern Cascadia. *Geophys Res Lett* 36: L07304. <https://doi.org/10.1029/2008GL036784>
- McCann GD, Wilts CH (1951) A mathematical analysis of the subsidence in the Long Beach – San Pedro area. California Institute of Technology, Pasadena, CA 117 p. (unpublished). <https://authors.library.caltech.edu/50357/>
- McGuire JJ, Segall P (2003) Imaging of aseismic fault slip transients recorded by dense geodetic networks. *Geophys J Int* 155:778–788. <https://doi.org/10.1111/j.1365-246X.2003.02022.x>
- McTigue DF (1987) Elastic stress and deformation near a finite spherical magma body; resolution of the point source paradox. *J Geophys Res* 92:12931–12940
- Metropolis N, Rosenbluth A, Rosenbluth M, Teller A, Teller E (1953) Equation of state calculations by fast computing machines. *J Chem Phys* 21:1087–1092
- Mogi K (1958) Relations between the eruptions of various volcanoes and the deformation of the ground surfaces around them. *Bull Earthq Res Inst U Tokyo* 36:99–134
- Moran SC (1994) Seismicity at Mount St. Helens, 1987–1992; evidence for repressurization of an active magmatic system. *J Geophys Res* 99:4341–4354
- Moran SC, Malone SD, Qamar AI, Thelen WA, Wright AK, Caplan-Auerbach J (2008) Seismicity associated with renewed dome building at Mount St. Helens, 2004–2005, chapter 2. In: Sherrod DR, Scott WE, Stauffer PH (eds) *A Volcano Rekindled: The Renewed Eruption of Mount St. Helens, 2004–2006*. U.S. Geological Survey. U.S. Geological survey professional paper 1750, Reston, pp 27–60
- Mosbrucker AR (2014) High-resolution digital elevation model of Mount St. Helens crater and upper North Fork Toutle River basin, Washington, based on an airborne lidar survey of September 2009. US geological survey data series 904. <https://doi.org/10.3133/ds904>
- Pavolonis MJ (2010) Advances in extracting cloud composition information from spaceborne infrared radiances: robust alternative to brightness temperatures. Part I: theory. *J Appl Meteorol Climatol* 49:1992–2012. <http://journals.ametsoc.org/doi/abs/10.1175/2010JAMC2433.1>
- Pavolonis MJ, Heidinger AK, Sieglaff J (2013) Automated retrievals of volcanic ash and dust cloud properties from upwelling infrared measurements. *J Geophys Res-Atmos* 118:1436–1458. <https://doi.org/10.1002/jgrd.50173>
- Pedersen R, Sigmundsson F, Masterlark T (2009) Rheologic controls on inter-rifting deformation of the Northern Volcanic Zone, Iceland. *Earth Planet Sci Lett* 281:14–26
- Perissin D, Wang T (2012) Repeat-pass SAR interferometry with partially coherent targets. *IEEE Trans Geosci Remote Sens* 50:271–280. <http://ieeexplore.ieee.org/ielx5/36/6117787/05978217.pdf?tp=&arnumber=5978217&isnumber=6117787>
- Pinel V, Poland MP, Hooper A (2014) Volcanology; lessons learned from synthetic aperture radar imagery. *J Volcanol Geotherm Res* 289(1):81–113. <https://doi.org/10.1016/j.jvolgeores.2014.10.010>
- Poland MP, Miklius A, Sutton AJ, Thornber CR (2012) A mantle-driven surge in magma supply to Kilauea Volcano during 2003–2007. *Nat Geosci* 5(4):295–300. (supplementary material, 16 p.). <https://doi.org/10.1038/ngeo1426>
- Poland MP, Miklius A, Montgomery-Brown EK (2014) Magma supply, storage, and transport at shield-stage Hawaiian volcanoes, chapter 5. In: Poland MP, Takahashi TJ, Landowski CM (eds) *Characteristics of*

- Hawaiian volcanoes. U.S. Department of the Interior, U.S. Geological Survey, Reston. U.S. Geological Survey Professional Paper 1801, p 429. https://pubs.usgs.gov/p1801/downloads/pp1801_Chap5_Poland.pdf
- Puskas C, Smith RB, Meertens CM, Chang WL (2007) Crustal deformation of the Yellowstone–Snake River Plain volcano-tectonic system; campaign and continuous GPS observations, 1987–2004, B03401. *J Geophys Res* 112. <https://doi.org/10.1029/2006JB004325>
- Riddick SN, Schmidt DA (2011) Time-dependent changes in volcanic inflation rate near Three Sisters, Oregon, revealed by InSAR. *Geochem Geophys Geosyst* 12: Q12005. <https://doi.org/10.1029/2011GC003826>
- Ring J (1963) The laser in astronomy. *New Sci* 344: 672–673
- Robinson JE (2014) Digital topographic data based on lidar survey of Mount Shasta Volcano, California, July–September 2010, U.S. Geological Survey data series, vol 852. <https://doi.org/10.3133/ds852>
- Ronchin E, Masterlark T, Molist JM, Saunders S, Tao W (2013) Solid modeling techniques to build 3D finite element models of volcanic systems; an example from the Rabaul Caldera system, Papua New Guinea. *Comput Geosci* 52:325–333. <http://www.sciencedirect.com/science/article/pii/S0098300412003354>
- Rymer H, Cassidy J, Locke CA, Murray JB (1995) Magma movements in Etna volcano associated with the major 1991–1993 lava eruption; evidence from gravity and deformation. *Bull Volcanol* 57:451–461
- Salvi S, Atzori S, Tolomei C, Allievi J, Ferretti A, Rocca F, Prati C, Stramondo S, Feillet N (2004) Inflation rate of the Colli Albani volcanic complex retrieved by the permanent scatterers SAR interferometry technique. *Geophys Res Lett* 31(12):12606. <https://doi.org/10.1029/2004GL020253>
- Scott WE (1987) Holocene rhyodacite eruptions on the flanks of South Sister volcano, Oregon. In: Fink JH (ed) The emplacement of silicic domes and lava flows, Geological Society of America special paper, vol 212. Geological Society of America, Boulder, pp 35–53
- Scott WE, Iverson RM, Schilling SP, Fischer BJ (2001) Volcano hazards in the Three Sisters region, Oregon. U.S. Geological survey open-file report 99–437, p 14
- Segall P, Matthews M (1997) Time dependent inversion of geodetic data. *J Geophys Res* 102:22,391–22,409. <https://doi.org/10.1029/97JB01795>
- Sherrod DR, Taylor EM, Ferns ML, Scott WE, Conrey RM, Smith GA (2004) Geologic map of the bend 30-by 60-minute quadrangle, Central Oregon. U.S. Geological survey miscellaneous field investigations map, I-2683, scale 1:100,000 scale, 48-page pamphlet
- Simons M, Agram PS, Philipposian B, Lin YN, Hetland EA, Fielding EJ, Yun S (2011) Exploiting ALOS observations as a guide to what will be possible with DESDynI, AGU Fall Meeting Abstracts, vol 22, 02. <http://adsabs.harvard.edu/abs/2011AGUFMIN22A..02S>
- Smith RB, Reilinger RE, Meertens CM, Hollis JR, Holdahl SR, Dzurisin D, Gross WK, Klingele EE (1989) What's moving at Yellowstone! – the 1987 crustal deformation survey from GPS, leveling, precision gravity and tri-lateralation. *EOS Trans Am Geophys Union* 70(8): 113–125
- Sparks RSJ (2003) Forecasting volcanic eruptions. *Earth Planet Sci Lett* 210:1–15. <http://www.sciencedirect.com/science/article/pii/S0012821X03001249>
- Stovall W, Cervelli P, Shelly D (2014) Investigating rapid uplift and subsidence near Norris, Yellowstone, during 2013–2014 [poster]. 2014 fall AGU meeting, V41B-4796
- Sutton AJ, Elias T (2014) One hundred volatile years of volcanic gas studies at the Hawaiian Volcano Observatory. Chapter 7. In: Poland MP, Takahashi TJ, Landowski CM (eds) Characteristics of Hawaiian volcanoes, U.S. Geological Survey professional paper, vol 1801. U.S. Department of the Interior, U.S. Geological Survey, Reston, pp 295–320
- Swanson DA (2008) Hawaiian oral tradition describes 400 years of volcanic activity at Kīlauea. *J Volcanol Geotherm Res* 176:427–431
- Symonds RB, Gerlach TM, Reed MH (2001) Magmatic gas scrubbing; implications for volcano monitoring. *J Volcanol Geotherm Res* 108:303–341
- Thelan WA, Crosson RS, Creager KC (2008) Absolute and relative locations of earthquakes at Mount St. Helens, Washington, using continuous data; implications for magmatic processes, chapter 4. In: Sherrod DR, Scott WE, Stauffer PH (eds) A volcano rekindled: the renewed eruption of Mount St. Helens, 2004–2006, U.S. Geological Survey professional paper, vol 1750. U.S. Geological Survey, Reston, pp 71–95
- Thompson RA, Schilling SP (2007) Photogrammetry. In: Dzurisin D (ed) Volcano deformation–geodetic monitoring techniques, chapter 6, Springer-Praxis books in geophysical sciences. Springer, Berlin, pp 195–221
- Thornber CR, Orr T, Heliker C, Hoblitt RP (2015) Petrologic testament to changes in shallow magma storage and transport during 30+ years of recharge and eruption at Kīlauea Volcano, Hawai'i, chapter 8. In: Carey RJ, Poland MP, Cayol V, Weis D (eds) Hawaiian volcanoes, from source to surface, AGU monograph. American Geophysical Union, Washington, DC, pp 147–188
- Tralli DM, Blom RG, Zlotnicki V, Donnellan A, Evans D (2005) Satellite remote sensing of earthquake, volcano, flood, landslide and coastal inundation hazards. *ISPRS J Photogramm Remote Sens* 59(4):185–198
- Trasatti E, Giunchi C, Bonafede M (2003) Effects of topography and rheological layering on ground deformation in volcanic regions. *J Volcanol Geotherm Res* 122:89–110
- Trasatti E, Cianetti S, Giunchi C, Bonafede M, Piana N, Agostinetti F, Casu F, Manzo M (2009) Bayesian source inference of the 1993–1997 deformation at Mount Etna (Italy) by numerical solutions. *Geophys J Int* 177:806–814. <https://doi.org/10.1111/j.1365-246X.2009.04093.x>
- Trasatti E, Bonafede M, Ferrari C, Giunchia C, Berrino G (2011) On deformation sources in volcanic areas; modeling the Campi Flegrei (Italy) 1982–84 unrest. *Earth Planet Sci Lett* 206(3–4):175–185. <https://doi.org/10.1016/j.epsl.2011.03.033>

- Vasco DW, Smith RB, Taylor CL (1990) Inversion for sources of crustal deformation and gravity change at the Yellowstone Caldera. *J Geophys Res* 95:19, 839–19,856
- Vasco DW, Puskas CM, Smith RB, Meertens CM (2007) Crustal deformation and source models of the Yellowstone volcanic field from geodetic data. *J Geophys Res* 112:B03401. <https://doi.org/10.1029/2006JB004325>
- Wallace PJ (2004) From mantle to atmosphere; magma degassing, explosive eruptions, and volcano volatile budgets. In: De Vivo B, Bodnar RJ (eds) *Melt inclusions in volcanic systems: methods, applications and problems*. Elsevier, Amsterdam, pp 105–127
- Werner C, Brantley S (2003) CO₂ emissions from the Yellowstone volcanic system. *Geochem Geophys Geosyst* 4(7):1061. <https://doi.org/10.1029/2002GC000473>
- Wicks C Jr, Thatcher W, Dzurisin D (1998) Migration of fluids beneath Yellowstone caldera inferred from satellite radar interferometry. *Science* 282:458–462
- Wicks CW Jr, Dzurisin D, Ingebritsen S, Thatcher W, Lu Z, Iverson J (2002) Magmatic activity beneath the quiescent Three Sisters volcanic center, Central Oregon Cascade Range, USA. *Geophys Res Lett* 29(7):26–1–26–4. <https://doi.org/10.1029/2001GL014205>
- Wicks CW, Thatcher W, Dzurisin D, Svarc J (2006) Uplift, thermal unrest, and magma intrusion at Yellowstone Caldera, observed with InSAR. *Nature* 440(7080): 72–75. <https://doi.org/10.1038/nature04507>
- Williams CA, Wadge G (1998) The effects of topography on magma chamber deformation models; application to Mt. Etna and radar interferometry. *Geophys Res Lett* 25:1549–1552
- Williams-Jones G, Rymer H (2002) Detecting volcanic eruption precursors; a new method using gravity and deformation measurements. *J Volcanol Geotherm Res* 113:379–389. <http://www.sciencedirect.com/science/article/pii/S0377027301002724>
- Wright CW, Brock J (2002) EAARL – a lidar for mapping shallow coral reefs and other coastal environments. In: Seventh international conference on remote sensing for marine and coastal environments, Miami, Fla., 20–22 May 2002, Proceedings: Ann Arbor, Michigan, Veridian International Conferences, CD-ROM
- Yamakawa N (1955) On the strain produced on a semi-infinite elastic solid by an interior source of stress. *J Seism Soc Japan* 8:84–98
- Yang X, Davis PM (1986) Deformation due to a rectangular tension crack in an elastic half-space. *Bull Seismol Soc Am* 76(3):865–881
- Academic Press, Waltham, p 376. <http://www.sciencedirect.com/science/book/9780123850485>
- Dzurisin D (2003) A comprehensive approach to monitoring volcano deformation as a window on the eruption cycle. *Rev Geophys* 41(1):1001. <https://doi.org/10.1029/2001RG000107>
- Dzurisin D (2007) *Volcano deformation—geodetic monitoring techniques*, Springer-Praxis books in geophysical sciences. Springer, Berlin, p 441
- Gelman A, Carlin JB, Stern HS, Rubin DB (2013) *Bayesian data analysis*, 3rd edn. CRC Press, Boca Raton/London/New York
- Gerya T (2010) *Introduction to numerical geodynamic modelling*. Cambridge University Press, p xvii + 345. <https://doi.org/10.1017/S0016756811000604>
- Hurwitz S, Lowenstern JB (2014) Dynamics of the Yellowstone hydrothermal system. *Rev Geophys* 51. <https://doi.org/10.1002/2014RG000452>
- Jaboyedoff M, Oppikofer T, Abellán A, Derron M-H, Loye A, Metzger R, Pedrazzini A (2012) Use of LIDAR in landslide investigations: a review. *Nat Hazards* 61(1):5–28. <https://doi.org/10.1007/s11069-010-9634-2>
- Lu Z, Dzurisin D (2014) *InSAR imaging of Aleutian volcanoes – monitoring a volcanic arc from space*, Springer-Praxis books, geophysical sciences, . ISBN 978-3-642-00347-9, p 388
- Massonnet D, Feigl KL (1998) Radar interferometry and its application to changes in the Earth's surface. *Rev Geophys* 36(4):441–500
- Menke W (2012) *Geophysical data analysis: discrete inverse theory*, 3rd edn. Elsevier Science and Technology, p 330. <http://www.sciencedirect.com/science/book/9780123971609>
- Parker RL (1994) *Geophysical inverse theory*, vol 400. Princeton University Press. ISBN 0-691-03634-9
- Reddy JN (2005) *An introduction to the finite element method*, 3rd edn. McGraw-Hill, p 912. ISBN 9780072466850
- Scales JA, Smith ML, Treitel S (2001) *Introductory geophysical inverse theory*. Samizdat Press, Golden, p 208. <http://citeseerx.ist.psu.edu/viewdoc/download?doi=10.1.1.87.9248&rep=rep1&type=pdf>
- Segall P (2010) *Earthquake and volcano deformation*. Princeton University Press, Princeton, p 432
- Siebert L, Simkin T, Kimberly P (2010) *Volcanoes of the world*, 3rd edn. Smithsonian Institution/University of California Press, Washington, DC/Berkeley, p 551
- Smithsonian Institution (2015) *Global volcanism program database*. National Museum of Natural History. <http://volcano.si.edu/#>
- Turcotte DL, Schubert G (2002) *Geodynamics*, 2nd edn. Cambridge University Press, p xv + 456. <https://doi.org/10.1017/S0022112002223708>

Books and Reviews

- Aster RC, Borchers B, Thurber CH (2013) *Parameter estimation and inverse problems*, 2nd revised edition.

Volcanoes in Iceland and Crustal Deformation Processes

Sigrún Hreinsdóttir
University of Iceland, Reykjavík, Iceland
GNS Science, Wellington, New Zealand

Article Outline

Glossary
Definition of the Subject
Introduction
Volcano Geodesy in Iceland
Deformation in Iceland
Volcano Deformation in Iceland
Bibliography

Glossary

Central Volcano Focal point of eruptive activity within a system. Often constructed by repeated eruptions from a central vent system that is maintained by a long-lived magmatic plumbing system.

Dike Magma body with thickness much smaller than its lateral extent. Formed by magma intruding into cracks.

Dry Tilt Dry-tilt measurements are high-precision optical leveling measurements performed on short lines oriented EW and NS or in a circle.

Fissure Swarm High-density subparallel volcanic fissures and tectonic fractures extending from a specific volcano.

Geodesy The study of the shape and area of the Earth, from large-scale variations that affect the rotation dynamics of the planet down to smaller length scales of earthquakes, landslides, etc.

GIA Glacial isostatic adjustment. The adjustment of the crust after an ice load is removed

from the surface. Here GIA is defined as the signal composed of the short-term elastic response and long-term viscoelastic adjustment due to the retreat of the ice caps in Iceland.

GNSS Global Navigation Satellite System. Satellite-based positioning systems, such as GPS, GLONASS, and GALILEO, with a network of satellites that transmit signals which can be used to infer three-dimensional positions.

GPS US Global Positioning System. A network of satellites that transmit signals which can be used to infer three-dimensional positions. The first GPS satellite was launched in 1978. The system had 24 satellites (full constellation) available by the end of 1993 and was declared fully operational in 1995.

InSAR Interferometric Synthetic Aperture Radar. The combination of Synthetic Aperture Radar imagery (generally acquired from airborne or satellite-based platforms) to infer ground deformation, digital elevation models, variations in atmospheric water vapor, etc.

Jökulhlaup Sudden glacial outburst flood.

Leveling Measurements of geodetic height relative to a reference surface. Optical leveling instrument are used to measure height between benchmarks in a line.

Magma Chamber Reservoir of molten rock (magma) in the crust.

Rift Linear zone where the Earth's crust is being pulled apart due to extensional tectonics.

Sill Tabular sheet intrusion, where magma has intruded between horizontal layers of rock.

Volcanic System Characterized by a central volcano, and often fissure swarm, a caldera, and high-temperature geothermal area associated with an upper crustal magma chamber.

Definition of the Subject

Crustal deformation processes associated with active volcanoes can vary from one volcano to

another and from one period to the next. Measurements of crustal deformation can provide important information on processes at depth and the state of a volcano. For a single reservoir system, pressure increases as magma accumulates in the reservoir, with gradual buildup of stress in the roof of the magma chamber. When a critical limit is reached, the wall of the magma chamber fails followed by intrusion or eruption. The preeruptive inflation and syn-eruptive deflation can be used to evaluate source depth and magma volume. The depth and the shape of the magma reservoir vary from one volcano to the next, and, at some volcanoes, the geodetic data suggest more than one chamber. The rate at which magma migrates into the reservoir can be episodic, and the effects of tectonic stresses and loads can change with time and evolve during volcanic events.

Many volcanoes are continuously deforming in response to internal process such as magma movements. Others are fairly quiet. Repeated geodetic measurements along with seismic monitoring are the key to better understanding volcanic processes and mitigating risk. Geodetic results from Icelandic volcanoes range from zero deformation or subsidence to continuously recharging magmatic systems (Dvorak and Dzurisin 1997; Massonnet and Sigmundsson 2000; Sigmundsson 2006; Segall 2010).

Introduction

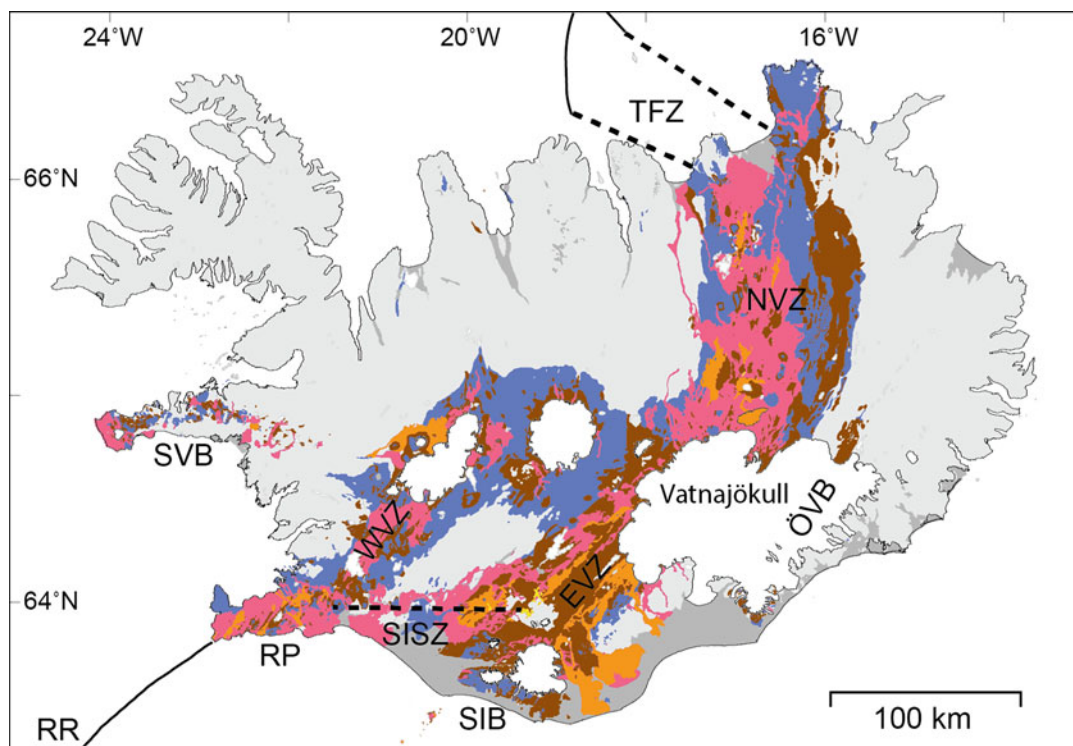
Iceland is located on the Mid-Atlantic Ridge where the rate of spreading between the North American and the Eurasian plate is about 18–19 mm/yr (DeMets et al. 2010). There are about 30 active volcanic systems in Iceland, and historically there has been on average an eruption every 4–5 years (Thordarson and Larsen 2007). The anomalously high volcanic activity and crustal thickness in Iceland has been explained by the existence of a mantle plume (Wolfe et al. 1997; Allen et al. 2002) interacting with the ridge. Iceland's existence above sea level makes it an ideal place to study crustal deformation and volcanic activity at a divergent plate boundary.

The Reykjanes Ridge (RR) emerges onshore at the southwest tip of the oblique spreading Reykjanes Peninsula (RP) (Hreinsdóttir et al. 2001; Keiding et al. 2008). At the Hengill triple junction, the RP connects to the West Volcanic Zone (WVZ) and the South Iceland Seismic Zone (SISZ). The SISZ transforms the main rift axis to the East Volcanic Zone (EVZ) (Sigmundsson et al. 1995), which is currently the volcanically most active region in Iceland (Thordarson and Larsen 2007). The full plate spreading in South Iceland is currently taken up over both the EVZ and the less active WVZ (La Femina et al. 2005; Geirsson et al. 2012). In north Iceland the active plate spreading takes place across the North Volcanic Zone (NVZ) (Pedersen et al. 2009; Drouin et al. 2014). The Tjörnes Fracture Zone (TFZ) shifts the rift system offshore to the Kolbeinsey Ridge (Rögnvaldsson et al. 1998; Metzger et al. 2011).

The Iceland plume has been active for the last 65 Ma (Saunders et al. 1997) and is currently centered beneath the EVZ, under the Vatnajökull ice cap. The oldest exposed rocks in Iceland are about 16 Ma old (Moorbath et al. 1968). The relative motion of the mantle plume and the spreading ridge has influenced the location of the plate boundary, causing a shift of the rift system east to Vatnajökull (Fig. 1; Sæmundsson 1978; Óskarsson et al. 1985; Einarsson 1991, 2008; Árnadóttir et al. 2009).

Three volcanic belts are located outside the main rift zone (Jakobsson 1972; Sæmundsson 1978; Óskarsson et al. 1985; Einarsson 2008). The Snæfellsnes Volcanic Belt in West Iceland is an old rift zone where volcanism was reactivated 2 Ma ago (Jóhannesson 1982). The South Iceland Volcanic Belt (SIB) is the southwestward propagating continuation of the EVZ but with little or no active rifting (Geirsson et al. 2012) and extends offshore to Vestmannaeyjar islands. The Örfajökull Volcanic Belt is mostly subglacial and east of the current rift axis. Three volcanic systems have been identified in the volcanic belt, but only Örfajökull volcano, the highest mountain in Iceland (2110 m), has verified eruptions during Holocene (Thordarson and Larsen 2007).

Thirty volcanic systems in Iceland have been active during the Holocene, of which 16 have



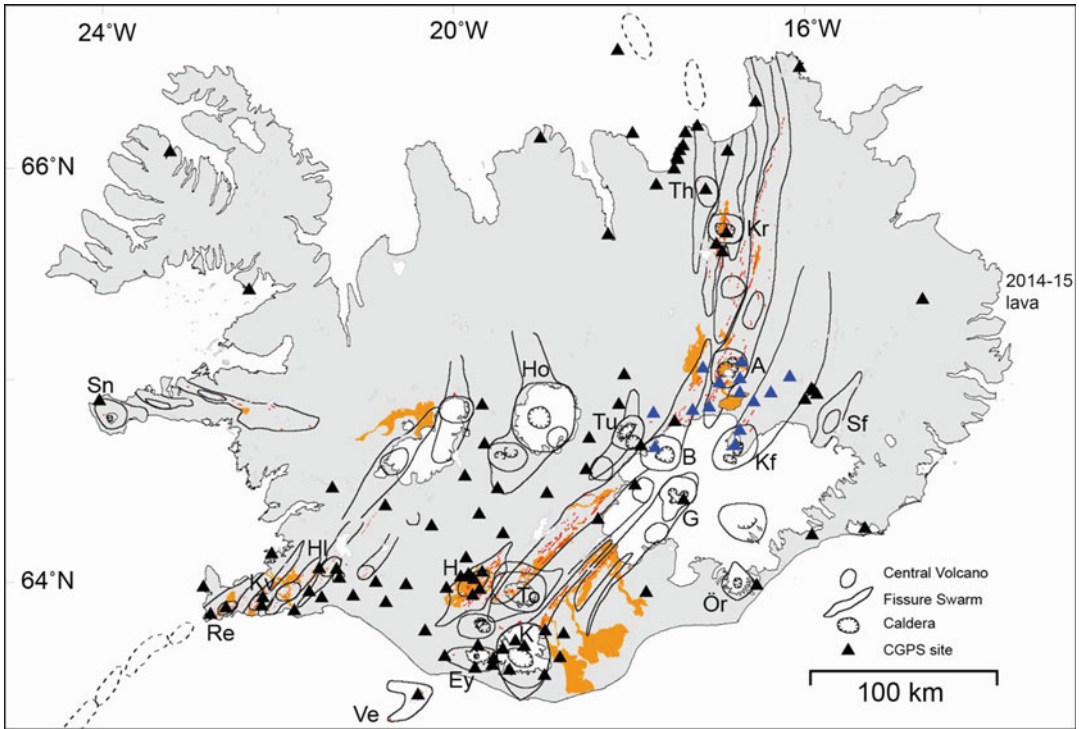
Volcanoes in Iceland and Crustal Deformation Processes, Fig. 1 Neo-volcanic zones in Iceland, defined by bedrock younger than 0.8 Ma. Interglacial lava (blue), hyaloclastite and pillow basalt (brown), younger than 0.8 Ma. Holocene prehistoric (pink) and historical (younger than 1100 years, orange) lava flows. Holocene sediments (dark gray). RR, Reykjanes Ridge; RP,

Reykjanes Peninsula; WVZ, West Volcanic Zone; EVZ, East Volcanic Zone; NVZ, North Volcanic Zone; SISZ, South Iceland Seismic Zone; TFZ, Tjörnes Fracture Zone; SIB, South Iceland Volcanic Belt; ÖVB, Öraefajökull Volcanic Belt; SVB, Snæfellsnes Volcanic Belt. (Jóhannesson and Saemundsson 2009a)

erupted in historical times (past 1100 years, Fig. 2) (Thordarson and Larsen 2007, and references therein). The Holocene volcanism is confined to 15–50 km-wide zones referred to as the neo-volcanic zones. Volcanism in Iceland is diverse, with basaltic, andesite, dacite, and rhyolite magmas erupted in effusive, explosive, and mixed eruptions during historical times. Phreatomagmatic basalt eruptions are common both for subglacial and submarine systems, but large volumes of basaltic lava have been erupted during rifting episodes, when magma intrudes into the fissure swarm (e.g., Sigmundsson et al. 2015; Thordarson and Self 1993; Sigurdardóttir et al. (2015); Einarsson 1991).

About 70% of all historical eruptions originate from the Grímsvötn, Hekla, Katla, and the

Bárdarbunga-Veidivötn volcanic systems. The most active volcanic system in Iceland is Grímsvötn, with over 60 documented eruptions in historical times (Larsen et al. 1998). The Grímsvötn fissure system extends about 90 km SW from the subglacial central volcano (Fig. 2). In 1783–1784 the Laki fires took place along a 27 km-long ice-free segment of the fissure swarm, producing about 14.7 km³ of lava (Thordarson and Self 1993). The Katla volcanic system in South Iceland has 21 confirmed historic eruptions, including the Eldgjá fissure eruption in 934 AD that produced the largest volume of lava erupted in historical times, about 20 km³ (Thordarson et al. 2001; Sigurdardóttir et al. 2015). The central volcano has a 110 km² and 700 m deep, ice-covered caldera. The other 20 historical Katla



Volcanoes in Iceland and Crustal Deformation Processes, Fig. 2 Active volcanic systems (Jóhannesson and Saemundsson 2009b) and CGPS sites (triangles) in Iceland. Holocene eruptive fissures (red) historical and lava flows (younger than 1100 years, orange) are shown. K,

Katla; G, Grímsvötn; B, Bárðarbunga; H, Hekla; Sn, Snæfellsjökull; Ö, Öraefajökull; Kr, Krafla; Sf, Snæfell; Th, Theistareykir; Kv, Kverkfjöll; Tu, Tungnafellsjökull; To, Torfajökull; Ve, Vestmannaeyjar; Re, Reykjanes; Hl, Hengill; Kk, Krisuvík; Ho, Hofsjökull; Ey, Eyjafjallajökull

eruptions were phreatomagmatic basalt eruptions within the caldera, generating jökulhlaups. Hekla volcanic system has 23 documented historical eruptions, the majority of which have been centered on the Hekla central volcano. The Hekla eruptions are mixed, where silica content of the first erupted material is dependent on the repose time since the previous eruption (Thordarson and Larsen 2007). The 190 km-long Bárðarbunga-Veidivötn system has erupted at least 27 times in the historical period, including both phreatomagmatic and fissure eruptions (Larsen and Gudmundsson 2014). On 31 August 2014, a fissure eruption started NE of the Bárðarbunga central volcano, after a 2-week seismic unrest and dike propagation over 45 km distance from the subglacial volcanic center (Sigmundsson et al. 2015). The eruption lasted until 27 February 2015 with $1.6 \pm 0.3 \text{ km}^3$ of

lava erupted (Gíslason et al. 2015), making it the largest eruption in Iceland since the 1783–1784 Laki eruption. During the eruption the ice-filled Bárðarbunga caldera subsided 65 m due to withdrawal of magma from a reservoir (Gudmundsson et al. 2016).

Volcano Geodesy in Iceland

Repeated geodetic measurements have been used to monitor and study volcano deformation processes in Iceland since the 1960s. In 1938, a triangulation network was set up across the NVZ centered at Krafla volcano, to test Wegener's continental drift hypothesis (Niemczyk 1943; Tryggvason 1984). The 100 km-long network was not measured again until 1965, using electronic distance measurements (EDM) and repeated

every few years for the next two decades alongside leveling profiles (Gerke 1974; Wendt et al. 1985). Leveling lines and optical leveling tilt stations (dry tilt) were established at several volcanoes in the 1960s to 1980s and measured regularly for monitoring purposes (e.g., Tryggvason 1968, 1989, 1994a, b, 1995; Sturkell et al. 2006). Watertube tilt meters were used at Hekla from 1968 to 1970 but were discontinued when it became apparent that dry-tilt measurements were of similar precision and easier to perform (Tryggvason 1994b). A 69 m-long watertube tilt meter was set up at the Krafla Power Plant in August 1976 to monitor deformation during the Krafla fires between 1975 and 1984. In 1977, continuously recording electronic tilt meters were installed in the Krafla region, which obtained higher precision and more frequent measurements (Tryggvason 1982). However, electronic meters are sensitive to temperature and drift slowly with time (Tryggvason 1995). In addition to more specialized geodetic equipment, a tape measure and creep meters were used during rifting episodes in Krafla to measure changes in fissure opening (e.g., Hauksson 1983). Lakes have been used as natural tilt meters, e.g., lake Öskjuvatn in Askja and lake Mývatn, SSE of Krafla volcano (Tryggvason 1987, 1989; Björnsson and Eysteinnsson 1998). In late 1979, a network of borehole strain meters to monitor Hekla was installed 15 to 45 km from the volcano (Linde et al. 1993).

During the past 25 years, Global Navigation Satellite System (GNSS) and Interferometric Synthetic Aperture Radar (InSAR) space geodetic techniques have been widely used to monitor and study volcanic deformation. GNSS geodetic measurements were first conducted in Iceland in 1986 when a 51 site, countrywide campaign network was established and measured using GPS data (US Global Positioning System) (Foulger et al. 1987). The measurements determine the location of a point on the ground with sub-centimeter level accuracy, allowing us to monitor the deformation of the surface with time. The first GNSS geodetic network to monitor volcanic activity was measured around Hekla volcano in 1989 (Sigmundsson et al. 1992). Several hundred geodetic benchmarks are now measured annually

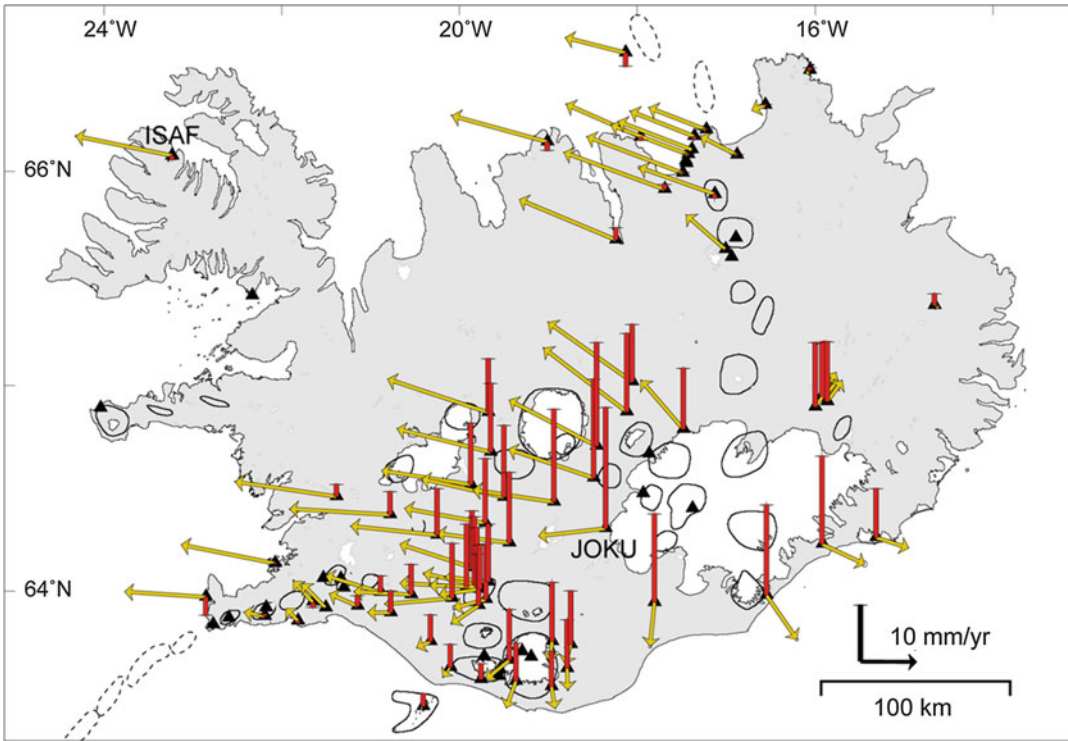
in Iceland. The first two continuous GNSS stations in Iceland were the International GNSS Service (IGS) stations Reykjavík (REYK), installed in 1995, and Höfn (HOFN), in 1997. In response to volcanic unrest, six GNSS stations were installed in 1999 and 2000 to monitor the Hengill, Eyjafjallajökull, and Katla volcanoes (Geirsson et al. 2006). The continuous GNSS network has grown rapidly in the last 10 years. In January 2014 there were 90 GNSS stations operating in Iceland, monitoring and measuring crustal deformation, the majority of which are located within the active neo-volcanic zone (Fig. 2). During periods of volcanic unrest, the GNSS network is augmented with semicontinuous stations (Hreinsdóttir et al. 2012; Sigmundsson et al. 2015).

InSAR measurements have been used to study volcanic deformation since 1992, after the launch of the first European Space Agencies (ESA) satellite, ERS1 (Massonnet et al. 1995; Massonnet and Sigmundsson 2000). By taking the phase difference between two SAR images, an interferogram can be constructed to reveal the surface deformation occurring in the time between two image acquisitions. When coherence is good, InSAR measurements provide a spatial coverage and resolution that can play an important role in understanding the nature of surface deformation (e.g., Ófeigsson et al. 2011a).

InSAR measurements to study volcanic processes in Iceland are mostly limited to the relatively ice-free volcanoes and conducted during the summer months. However, during the 2014 Bárðarbunga eruption, InSAR data were used along with airplane radar profiling and GNSS measurements to monitor the rapid subsidence of the ice-covered Bárðarbunga caldera (Sigmundsson et al. 2015; Gudmundsson et al. 2016). InSAR techniques have also been used to study surface processes such as ice melting due to volcano-glacier interaction (Björnsson et al. 2001).

Deformation in Iceland

Crustal deformation in Iceland is caused by multiple sources. In order to study and monitor



Volcanoes in Iceland and Crustal Deformation Processes, Fig. 3 Crustal deformation in the ITRF08 Eurasian reference frame at CGPS stations in Iceland, red bars showing vertical velocities and yellow arrows showing horizontal. The velocities are based on continuous GPS

measurements from 1996 to 2014.0 excluding data affected by volcanic unrest, post-seismic deformation, and geothermal exploitation. Velocities are only shown for sites where time series exceed 2.5 years

volcanic deformation, it is important to understand the background signals.

Iceland is located at the divergent plate boundary between the North American and Eurasian plates. A GNSS station at Ísafjörður (ISAF), at the north-west tip of Iceland, is moving at a steady rate of 17.2 ± 0.1 mm/yr in direction ENE relative to the stable ITRF08 Eurasian reference frame (Fig. 3; Altamimi et al. 2012). The deformation in the neo-volcanic zone is episodic with rifting events in the fissure swarms and earthquakes in the seismic zones separated by long periods of relative quiescence (Sigmundsson 2006; Sigmundsson et al. 2018, and reference therein).

Glaciers and ice caps cover about 11% of the country (Björnsson and Pálsson 2008). The glaciers are melting and retreating, causing the crust to rebound. The Earth's elastic and viscoelastic

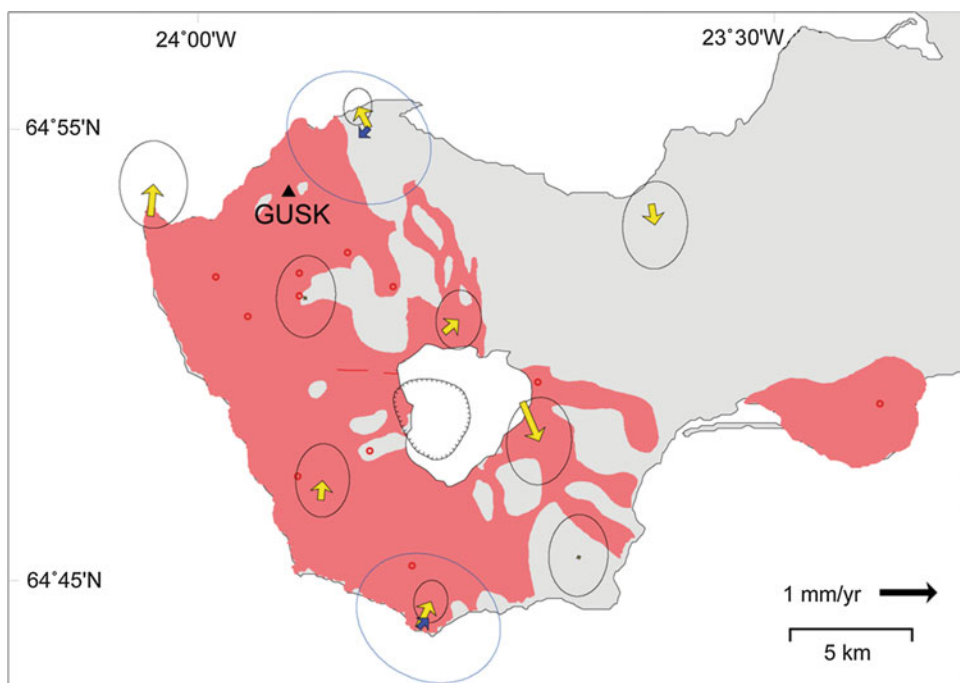
response to ice retreat results in uplift rates currently exceeding 35 mm/yr in central Iceland (Compton et al. 2015; Fridriksdóttir 2014; Fig. 3) and horizontal movement away from the ice caps (Árnadóttir et al. 2009; Geirsson et al. 2010; Auriac et al. 2013). Annual load changes due to snow and ice (Grapenthin et al. 2006; Drouin et al. 2016; Compton et al. 2017) cause an elastic response of the crust. Other load changes such as the draining of lakes, surging glaciers (Auriac et al. 2014), or the embedment of new lava flows (e.g., Grapenthin et al. 2010) can result in a significant deformation signal. In addition, young lava flows cool and contract, which can be observed in InSAR images (e.g., Sigmundsson et al. 1997a; Ófeigsson et al. 2011a; Wittmann et al. 2017).

Geothermal utilization of high-temperature geothermal field associated with a volcanic

system can result in localized subsidence signal, resulting from pressure change and/or temperature decrease in the geothermal reservoir. Several geothermal power plants are operating in Iceland both extracting geothermal fluids and reinjecting wastewater. In SW Iceland four geothermal power plants are currently in operation, two on Reykjanes Peninsula (Reykjanes and Svartsengi) and two in Hengill (Hellisheiði and Nesjavellir). In North Iceland three geothermal power plants are in operation, at Krafla, Theistareykir, and Bjarnarflag. Deformation related to these power plant has been monitored using leveling, GNSS, and InSAR measurements (e.g., Keiding et al. 2010; Michalczevska et al. 2014; Juncu et al. 2017; Drouin et al. 2017; Parks et al. 2018; Juncu et al. 2018).

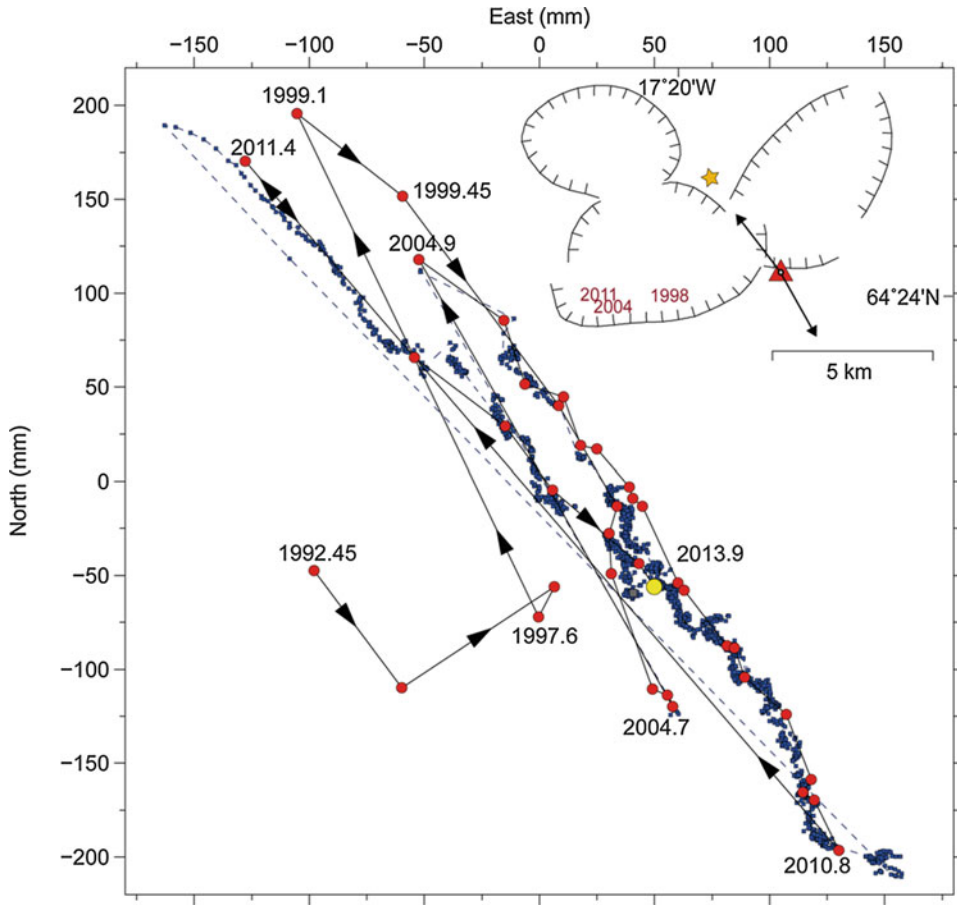
Volcano Deformation in Iceland

Styles of volcano deformation in Iceland vary from one volcano to the other and from one period to the next. Some volcanoes are very active, with continuous deformation relating to pressure changes at depth (e.g., Sigmundsson et al. 2014). Others show little signs of life, even over long periods. Snæfellsjökull, a stratovolcano in West Iceland, has erupted several times during Holocene, but no historical volcanic activity has been documented (Thordarson and Larsen 2007). GPS measurements were conducted around the volcano in 1993, 2004, and 2013, and in 2012 a GNSS station was installed 12 km NNE of the volcano center (Fig. 4). The measurements show no significant deformation of the volcano (Kristinsson 2014).



Volcanoes in Iceland and Crustal Deformation Processes, Fig. 4 Horizontal velocities with 95% uncertainty ellipses around Snæfellsjökull volcano from 2004.5 to 2013.5 (yellow, Kristinsson 2014) and 1993.5 to 2004.5 (blue, Árnadóttir et al. 2009). The average velocity in the

region has been subtracted from the velocity estimates. Holocene lava flows (pink), craters, and eruptive fissures (red) are shown in addition to the GNSS station GUSK operating since 2012



Volcanoes in Iceland and Crustal Deformation Processes, Fig. 5 Horizontal movement of the GPS site GRIM (red) and GNSS station GFUM (blue) at Mt. Grímsfjall (red triangle), showing inflation/deflation signals due to pressure changes at a shallow magma

chamber at the center of the caldera complex (star) (updated from Hreinsdóttir et al. 2012). The yellow dot marks 2013.9. The locations of the 1998, 2004, and 2011 eruptions are shown in the inset (upper right corner)

Over 20 confirmed eruptions have occurred in Iceland since 1970, including 9 during the Krafla fires. Measured crustal deformation leading up to an eruption has varied from a steady inflation of a crustal magma chamber over several years (e.g., Ófeigsson et al. 2011a) to a more complex episodic movement of magma into dikes and sills for a few weeks to years prior to an eruption (Sigmundsson et al. 2010, 2015; Hjaltadóttir et al. 2015). Co-eruptive deformation often involves more than one source, with the formation of feeder dikes and the depletion of magma source(s) at depth (e.g., Sturkell et al. 2013; Hreinsdóttir et al. 2014).

Grímsvötn

Grímsvötn volcano is the most active volcano in Iceland (e.g., Óladóttir et al. 2011). Its location, beneath the Vatnajökull ice cap, makes deformation studies challenging, limiting measurements to a single nunatak, Mt. Grímsfjall, at the southern caldera rim (Fig. 5). GPS measurements spanning over three eruptive cycles show uplift and displacement away from the caldera, at a rate of a few cm each year, interrupted by a sudden co-eruptive subsidence and movement toward the caldera (Fig. 5; Sigmundsson et al. 2014; Sturkell et al. 2003a, 2006). In addition, the site is uplifting in response to the melting of the Vatnajökull ice cap at

a rate exceeding 20 mm/yr (Geirsson et al. 2010; Fridriksdóttir 2014; Compton et al. 2015). Measurements using co-located GNSS and tilt station during the 2011 Grímsvötn eruption suggest a pressure drop in a magma chamber at 1.7 $\pm$ 0.2 km depth (Hreinsdóttir et al. 2014). The initiation of the pressure drop preceded the eruption by an hour and reached maximum within the first 48 h of the eruption. The long-term GPS time series, showing rapid exponentially decaying post-eruptive uplift for a few years followed by approximately linear uplift rates, have been interpreted in terms of a two-magma reservoir system due to a pressure readjustment between reservoirs and a constant inflow of magma into the deeper reservoir (Reverso et al. 2014).

Hekla

Hekla, a ridge-shaped stratovolcano in South Iceland, has 18 confirmed eruptions since 1104 AD, with 1 to 3 eruptions per century (Thordarson and Larsen 2007). In the past 50 years, the volcano has erupted at a regular and more frequent rate, with relatively small eruptions in 1970, 1980–1981, 1991, and 2000. Hekla eruptions usually start with a short Plinian or subplinian phase followed by lava fountaining and effusive activity (Thordarson and Larsen 2007). The longer the repose time, the larger and more silica rich are the eruptions (Thorarinsson 1967). This suggests a reservoir where magma can accumulate and evolve with time.

Hekla is one of the most closely geodetically monitored volcanoes in Iceland. In addition to regular GPS campaigns and measurements of dry-tilt stations, a network of seven GNSS stations is operated on the volcano (Geirsson et al. 2010), and five borehole strain meters (Linde et al. 1993) are located at 15–45 km distance. Very few inter-eruptive earthquakes have been detected at Hekla (e.g., Soosalu and Einarsson 1997) suggesting that the magma reservoir at Hekla volcano is deeply seated. In 1991 and 2000 a swarm of small earthquakes started 30 and 79 min prior to the eruptions, respectively (Linde et al. 1993; Soosalu et al. 2005; Einarsson 2018). The formation of a shallow feeder dike was detected with the strain meter about half an hour prior to the 2000 eruption (Sturkell et al. 2013; Figs. 6 and 7).

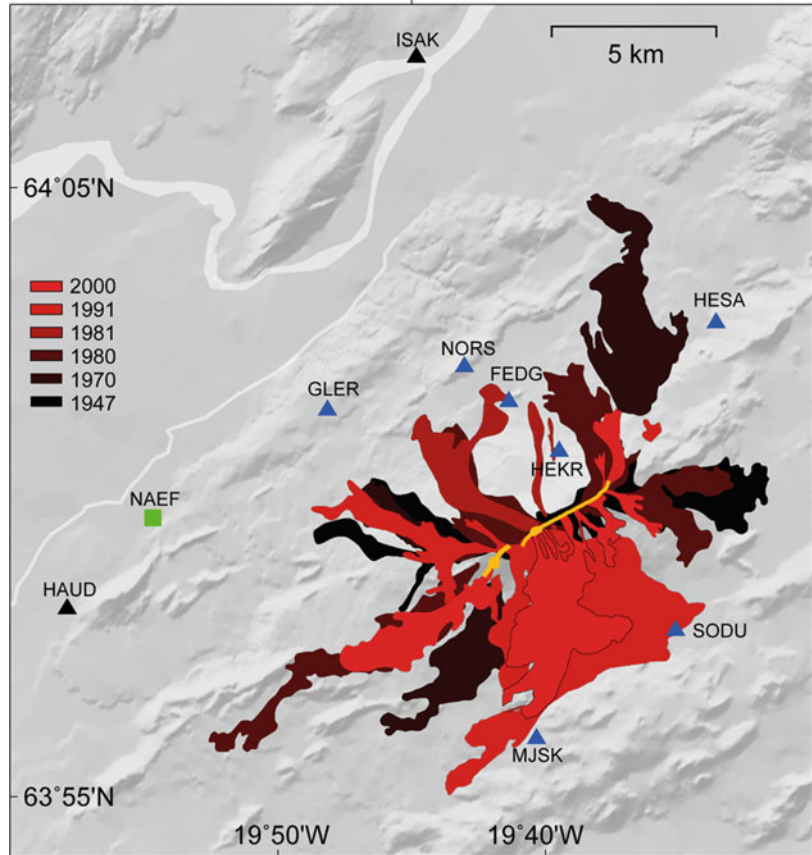
Geodetic measurements around Hekla volcano during the last three eruptive cycles have shown a steady rate of inflation between eruptions (Kjartansson and Gronvold 1983; Tryggvason 1994b; Ófeigsson et al. 2011a; Geirsson et al. 2012) and rapid co-eruptive deflation (Tryggvason 1994b; Sigmundsson et al. 1992; Ófeigsson et al. 2011a; Sturkell et al. 2013). The data have been used to constrain the depth of the magma chamber beneath Hekla resulting in a wide range of estimates, from 5–6 km (Tryggvason 1994b) to 22–29 km depth (Geirsson et al. 2012). Co-eruptive strain measurements indicate a reservoir at around 10 km depth (Sturkell et al. 2013), but recent GPS and InSAR data suggest a more deeply seated magma chamber (over 14 km depth) (Geirsson et al. 2012; Ófeigsson et al. 2011a), in agreement with seismic observations (Soosalu and Einarsson 2004). Both co-eruptive InSAR and strain meter data indicate a shallow feeding dike that does not extend all the way to the reservoir (Sturkell et al. 2013). During early surveys of the volcano, far-field and near-field tilt measurements sometimes gave contradicting results (e.g., Tryggvason 1994b). This discrepancy could possibly have originated from the shallow dike feeding the eruption. In addition, a localized subsidence signal is observed in InSAR images around recent lava flows around the summit of Hekla (Ófeigsson et al. 2011a). The subsidence is caused by both thermal contraction of the lava itself and loading effects (Ófeigsson et al. 2011a, Grapenthin et al. 2010; Wittmann et al. 2017). The extent of the signal is limited, but it can bias estimates of source parameters.

Katla

The Katla volcano lies under the Mýrdalsjökull ice cap in South Iceland. It has 20 confirmed phreatomagmatic basalt eruptions within the ice-filled caldera over the past 1100 years (Thordarson and Larsen 2007). A low-velocity anomaly and a clear S-wave shadow beneath the Katla caldera have been interpreted in terms of a magma chamber at shallow depth (Gudmundsson et al. 1994). Katla is seismically the most active volcano in Iceland, with periods of elevated unrest since 1952, most recently in 1999–2004 (Sturkell

Volcanoes in Iceland and Crustal Deformation Processes, Fig. 6

CGNSS stations around Hekla volcano (triangles) and dry-tilt station NAEF (square). Recent lava flows (colored black to red) and the eruptive fissure in 2000 (yellow) are also shown

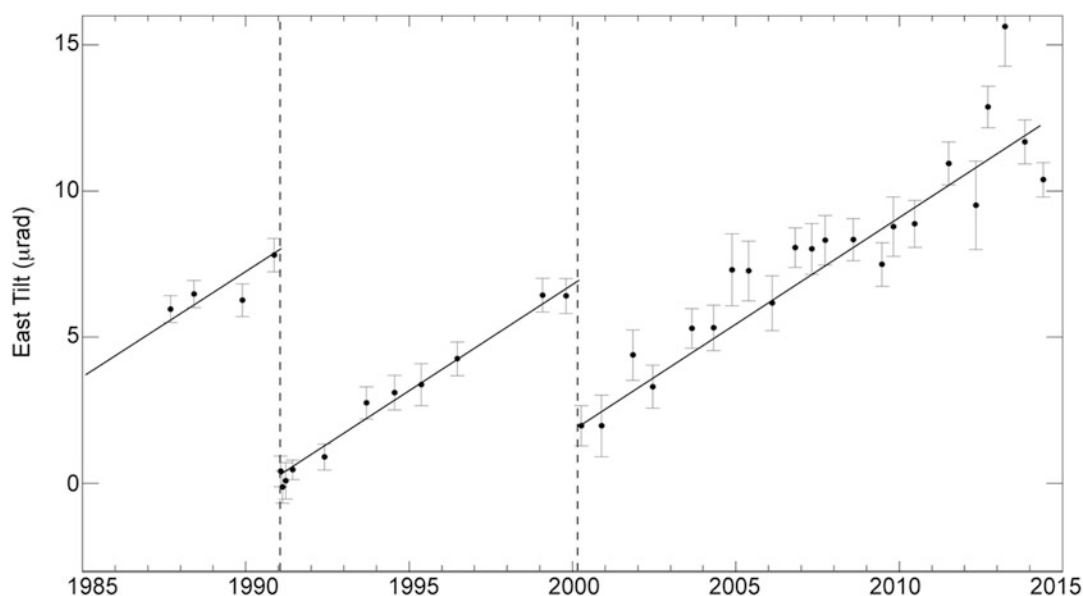


et al. 2008b) and 2011–2012 (Ófeigsson et al. 2011b). The seismic activity also shows annual variability, with activity peaking in the fall, possibly related to elevated fluid pressure and reduced ice load (Einarsson and Brandsdóttir 2000). Snow thickness variations of up to 6 m have also been considered a triggering factor for seasonal seismicity as well as eruptions (Albino et al. 2010). Interestingly, documented Katla eruptions tend to peak in the late summer and fall, with no documented eruption starting from December through April (Thordarson and Larsen 2007).

Since the 1100s, one to three Katla eruptions have occurred per century, each accompanied by a jökulhlaup. In contrast to observations at Hekla volcano, the magnitudes of Katla eruptions do not appear to correlate with the length of the preceding repose interval. However, longer eruption intervals tend to follow large eruptions (Eliasson et al. 2006). The most recent Katla eruption took place in 1918. It was a big eruption, accompanied by a

massive glacial outburst flood (e.g., Thordarson and Larsen 2007 and reference therein). Three smaller events, increased seismic activity accompanied by a small glacial outburst floods, occurred in 1955, 1999, and 2011. It has been suggested that these floods were caused by small subglacial eruptions or shallow magma intrusions under the glacier (Einarsson and Brandsdóttir 2000; Ófeigsson et al. 2011b).

Geodetic studies of Katla volcano are limited to the volcano flank, outside the Mýrdalsjökull ice cap, and nunataks at the caldera rim. In 1966, three dry-tilt stations were established east of Mýrdalsjökull glacier to monitor the volcano. Measurements from 1967 to 1971 showed repeated uplift and subsidence that correlated with snow load variations on the glacier (Tryggvason 1973). GPS campaign measurements outside the glacier have shown steady uplift and deformation away from the center of the volcano since measurements started in 1992. Two GNSS stations have been



Volcanoes in Iceland and Crustal Deformation Processes, Fig. 7 Measurements at the dry-tilt station Næfurholt (NAEF in Fig. 6), 11 km west of the Hekla summit. Dashed lines mark Hekla eruptions. The station

shows inter-eruptive inflation (tilt of about $0.75 \mu\text{rad/yr}$) and syn-eruptive deflation of the volcano. (Modified and updated from Sturkell et al. 2013, courtesy Erik Sturkell 2014)

operating at the southern flank of the volcano since 1999, and another was set up over an existing benchmark on the western flank of the volcano in 2006. The sites show a combination of annual variations due to load changes (Grapenthin et al. 2006; Pinel et al. 2007) and a steady uplift signal. To evaluate how much of the observed uplift was due to glacial retreat in the region rather than of volcanic origin, measurements of glacial ice thickness variations were used to predict the response of the crust to the glacier thinning (Pinel et al. 2007; Sturkell et al. 2008b). The models suggested that the observed horizontal measurements around Katla were too large compared to the vertical data to be caused by rebound only, thus concluding that the data were consistent with magma accumulation beneath the volcano. The study, however, only considered the contribution of the Mýrdalsjökull ice cap in the calculations. InSAR time series spanning 2003 to 2009 from South Iceland show no sign of inflation around Katla volcano after being corrected for GIA model calculations (Spaans 2011; Spaans et al. 2015).

Campaign GPS data from nunataks within the Mýrdalsjökull glacier suggested rapid inflation during the 2000 to 2004 period (Sturkell et al.

2008b), which along with seismic unrest was considered a sign that Katla was preparing for an eruption. In 2010, continuous GNSS stations were set up at the nunataks, revealing a complex deformation signal presumably originating from within the caldera (Ófeigsson et al. 2011b; Spaans et al. 2015). The signal is annual with up to 60 mm variations in the horizontal and 70 mm in the vertical. Any measurement at the nunataks would therefore have been highly dependent on their timing.

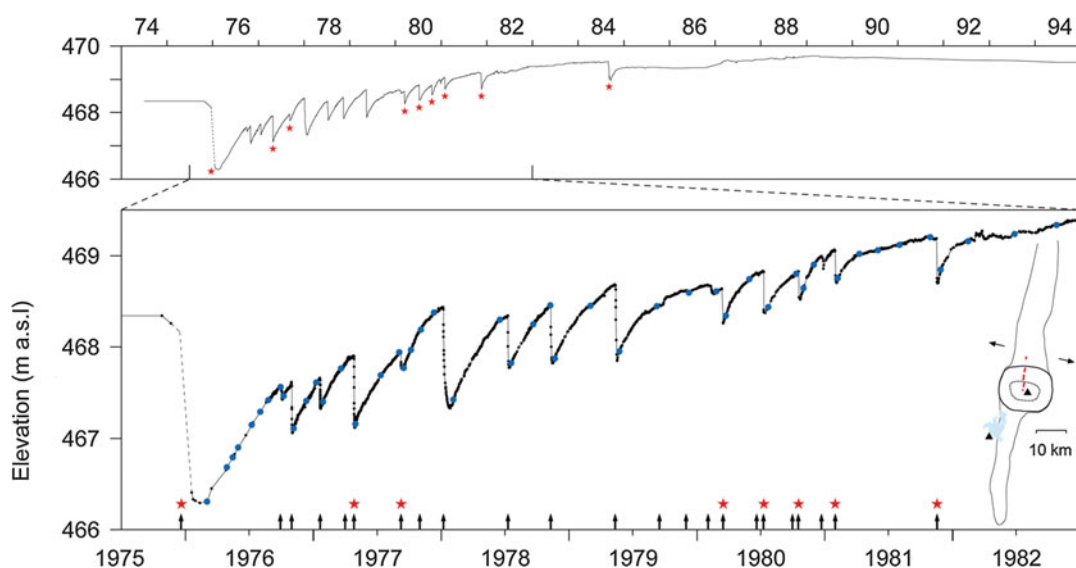
Krafla

The Krafla volcanic system in the NVZ has a central volcano and a shallow magma chamber below a 10 km-wide caldera (Einarsson 1978; Brandsdóttir et al. 1997). A 90 km-long fissure swarm extends north and south from the central volcano. The volcanic activity at Krafla is characterized by rifting episodes separated by long periods of quiescence. Two rifting episodes have occurred in Krafla during historical times, the Mývatn “fires” of 1724–1729 and the Krafla “fires” of 1975–1984. The term “fires” refer to volcano-tectonic episodes during which short periods of high eruptive activity, with rifting and

faulting along the plate boundary, are separated by more quiet periods. The 1975–1984 Krafla fires were well documented. EDM measurements and leveling had been repeated every 2–5 years from 1965 to 1975. The data suggest a contraction of the volcano from 1965 to 1970 but slight expansion from 1970 to 1975 accompanied by increased seismic activity (Wendt et al. 1985; Björnsson et al. 1977). From December 1975 to September 1984, about 21 rifting events took place in the Krafla volcanic system, with 9 leading to surface eruptions (Fig. 8). The first event began in December 1975 and lasted for a few weeks (Björnsson et al. 1977). It started with intense seismic activity and volcanic tremor and then an eruption breaking out in the center of the caldera. The earthquakes propagated northward with about a 60 km-long segment of the fissure swarm affected with widening of fissures and graben formation. About 36 earthquakes of magnitude 4.5 and larger occurred during the rifting event, the largest earthquake magnitude 6.3. During the rifting event, the caldera floor subsided over 2 m (Fig. 8). Following the event, seismic activity in the fissure swarm

gradually decreased, and in March 1976 the volcano started inflating, with maximum uplift near the center of the caldera (Björnsson et al. 1977).

Leveling and tilt measurements at Krafla volcano provided important information both for monitoring and forecasting rifting events during the Krafla fires and to study the dynamics of rifting processes (e.g., Buck et al. 2006; Wright et al. 2012). Continuous uplift with gradually increasing seismic activity within the caldera for a few weeks to months was followed by a rapid subsidence and volcanic tremor from a couple of hours to several days, often associated with an earthquake swarm in the fissure swarm and increased geothermal activity (e.g., Björnsson et al. 1979; Einarsson 1991). The initial rate of uplift of the Krafla caldera following each rifting event was about 6 mm/day but then slowed down with earthquake activity increasing steadily. The observations are consistent with a magma flow rate of about $5 \text{ m}^3/\text{s}$ into a 3 km deep magma chamber (Björnsson et al. 1979) from a deeper reservoir (Tryggvason 1986; Árnadóttir et al. 1998). When a critical pressure was reached,



Volcanoes in Iceland and Crustal Deformation Processes, Fig. 8 Elevation changes from 1974 to 1995 at a benchmark FM5596, 1 km SE of the center of the Krafla caldera. The line shows interpolated values based on leveling (blue circles) and tilt measurements (black dots). The leveling was done relative to a benchmark at the southern

shore of Lake Mývatn, 20 km away from the caldera (FM6414). Arrows indicate rifting events, red stars eruptions and triangles show geodetic sites FM5596 and FM6414. (Data from Björnsson and Eysteinnsson 1998 and references therein)

the magma extruded laterally out of the magma chamber and into the fissure swarm, where widening of fissures and subsidence was observed. A maximum accumulated widening of about 8–9 m was measured from 1975 to 1984 across the central part of the fissure system (Tryggvason 1984, 1994a), corresponding to 500 years of long-term plate spreading.

During 1980–1981, four eruptions took place along fissures extending 10 km north of the caldera. The last rifting event took place in September 1984 with the largest eruption of the sequence, an 8.5 km-long fissure eruption with 0.11 km³ of lava (dense rock equivalent) erupted (Rossi 1997). During the Krafla fires, only a small portion of the estimated magma leaving the shallow magma chamber erupted at the surface. After the eruption in 1984 ended, inflation continued for about 5 years (Tryggvason et al. 1994a). Since 1989 a steady subsidence has been observed in the region (Fig. 8).

InSAR and GPS measurements, along with leveling and tilt measurements, have been conducted regularly around Krafla since the end of the Krafla fires (e.g., Foulger et al. 1992; Sigmundsson et al. 1997a; Sturkell et al. 2008a; de Zeeuw-van Dalfsen et al. 2004). The data show subsidence along the fissure swarm and continued deflation of the magma chamber, with rate decreasing with time. In addition localized subsidence signal around boreholes has been observed at the Krafla power station (Drouin et al. 2017). A broad signal of uplift seen in InSAR images spanning 1993–1999 has been interpreted as a deep-seated magma accumulation at the volcano (de Zeeuw-van Dalfsen et al. 2004).

Bárdarbunga

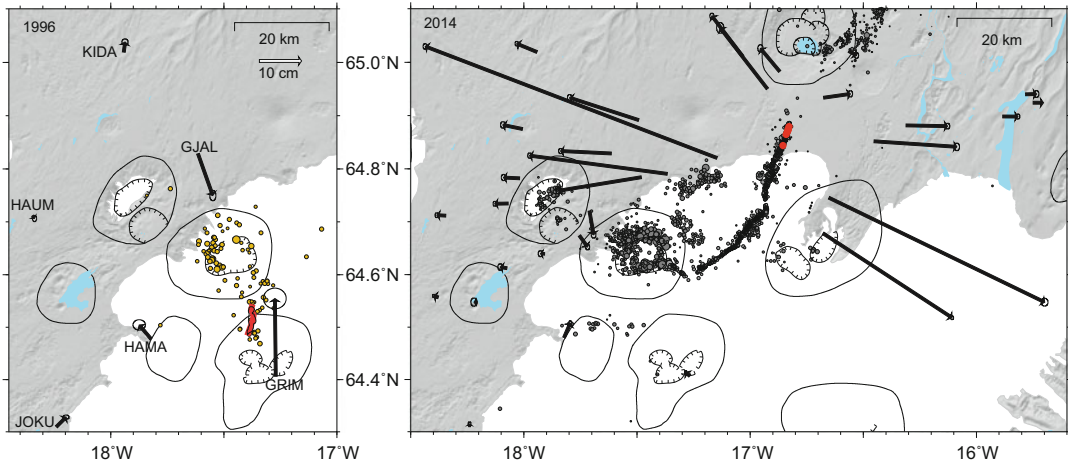
Bárdarbunga volcano in central Iceland lies beneath the northwestern part of the Vatnajökull ice cap. It rises over 2000 meters above sea level and has an 80 km² ice-filled caldera. The Bárdarbunga-Veidivötn fissure swarm extends 115 km SSW and 55 km NNE from the central volcano (Thordarson and Larsen 2007).

On 29 September 1996 at 10:48 UTC, an Mw 5.6 event occurred at the northern rim of Bárdarbunga caldera, followed by an intense earthquake swarm (Einarsson et al. 1997; Pagli

et al. 2007). The earthquakes migrated 20 km south toward the Grímsvötn volcano over the next 24 h. In the evening of 30 September, a subglacial eruption, the Gjálp eruption, began between the two volcanic centers (Einarsson et al. 1997; Gudmundsson et al. 2004). The eruption took place on a 6 km-long, NS trending fissure, initially below 550 to 750 m thick ice (Gudmundsson et al. 2004). On 2 October the eruption melted an opening through the ice resulting in a phreatomagmatic eruption that lasted until 13 October.

An estimated 0.45 km³ of magma (dense rock equivalent) erupted during the 2-week Gjálp eruption (Gudmundsson et al. 2004). Analysis of the isotopic composition of the erupted material showed that the Gjálp magma did not originate from the Bárdarbunga volcano but rather the Grímsvötn volcanic system (Sigmarsson et al. 2000). Limited geodetic data are available for the region due to the ice cover, but campaign GPS measurements in the following summer showed that a site located NNE of Bárdarbunga (GJAL) had moved toward Bárdarbunga, suggesting deflation of the volcano (Hreinsdóttir et al. 1998; Pagli et al. 2007, Fig. 9). InSAR data outside the ice cap also show significant deformation around the volcano which could be explained by a deflating source at around 10 km depth beneath Bárdarbunga. However, the deflation had to have taken place after 6 October (Pagli et al. 2007). Syn-eruptive interferograms that span the first few days of the eruption show no significant deflation. Rather, the majority of the signal is consistent with dike at the Bárdarbunga caldera rim. The InSAR data were not sensitive to possible dike events south of Bárdarbunga (Pagli et al. 2007).

On 16 August 2014, an intense seismic swarm began at Bárdarbunga volcano (Sigmundsson et al. 2015). The earthquakes migrated from the Bárdarbunga caldera in segments with variable strike over 45 km distance for nearly 2 weeks. GPS observations show both deflation of the volcano and displacement consistent with the growth of the dike (Fig. 9). On 24 August a series of M5+ earthquakes started around the Bárdarbunga caldera rim. Airborne radar profiling data and 1-day SAR interferograms over the



Volcanoes in Iceland and Crustal Deformation Processes, Fig. 9 Deformation around Bárðarbunga volcano during eruptive episodes in 1996 and 2015. The left panel shows deformation related to the 1996 Gjalp eruption (updated and modified from Hreinsdóttir et al. 1998) with earthquakes in yellow and eruption site in red (Einarsson

et al. 1997). The right panel shows deformation from 16 August to 6 September 2014 during the dike intrusion and first days of the Holuhraun eruption (Sigmundsson et al. 2015). Earthquakes are shown in orange (dike) and gray and eruption site in red

ice surface revealed up to about 1 m/day subsidence of the caldera floor.

The propagation of the dike was monitored with seismic data, continuous GPS measurements, and InSAR data (Sigmundsson et al. 2015). Significant extension was observed over 100 km distance ESE and WNW of the volcano (Ófeigsson et al. 2015). On 29 August a minor fissure erupted for a few hours in Holuhraun, northeast of Bárðarbunga. On 31 August the eruption started again with a 1.5 km-long fissure erupting. Widening across the dike continued at decaying rate for the first week of September, but GNSS stations continued to respond to the deflating Bárðarbunga volcano (Sigmundsson et al. 2015; Gudmundsson et al. 2016). The eruption ended on the 27 February 2015 with $1.6 \pm 0.3 \text{ km}^3$ of lava erupted (Gíslason et al. 2015). During the Bárðarbunga unrest and in the months following the eruption, several new GNSS stations were installed to monitor the volcano (Ófeigsson et al. 2015). Following the eruption InSAR and GPS data reveal post-rifting relaxation and isostatic adjustment to the new lava flow (Grapenthin et al. 2018).

The Gjalp 1996 and Holuhraun 2014 eruptions were both initiated by seismic activity at the

Bárðarbunga caldera followed by a complex dike propagation and rifting (Einarsson et al. 1997; Pagli et al. 2007; Sigmundsson et al. 2015; Gudmundsson et al. 2016). The events involved neighboring volcanic systems, most significantly the 1996 Gjalp magma having the geochemical signature of Grímsvötn volcanic system. During both 1996 and 2014, triggered seismicity was observed around Tungnafellsjökull volcano (Fig. 9), and InSAR data from 1996 reveal triggered slip on faults up to 30 km distance from Bárðarbunga (Pagli et al. 2007; Parks et al. 2017).

Eyjafjallajökull

Eyjafjallajökull in South Iceland is an ice-capped volcano with a summit crater, located outside the main rift zone. Only three historical eruptions are known at Eyjafjallajökull prior to 2010, a fissure eruption in 920 and summit eruptions in 1612 and 1821–1823 (Larsen et al. 1999; Óskarsson 2009). On 20 March 2010, a fissure eruption started on the east flank of the volcano followed by an explosive summit eruption on 14 April.

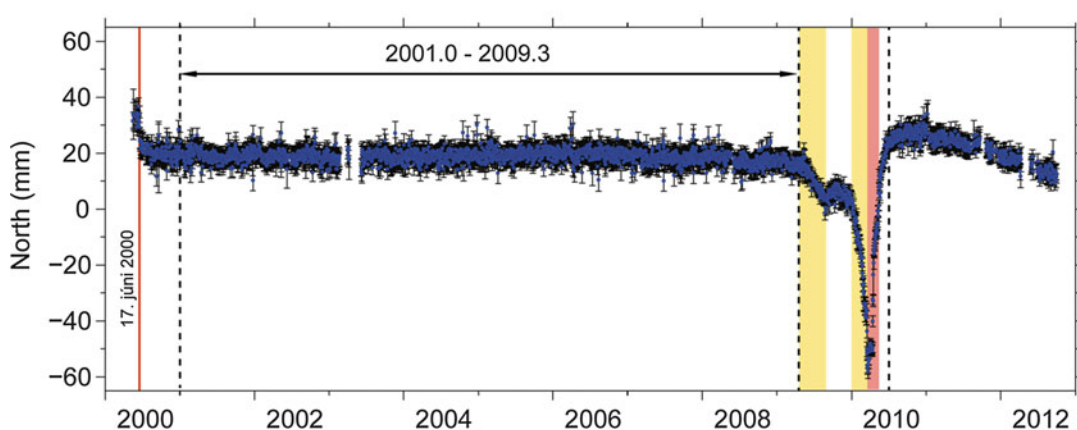
The 2010 eruptions were preceded by 18 years of intermittent unrest. In 1994 and 1999 sill

intrusions formed at 4.5–6.5 km depth beneath Eyjafjallajökull (Pedersen and Sigmundsson 2004, 2006; Hooper 2008; Sturkell et al. 2003; Hjaltadóttir et al. 2015). The intrusions were mapped by InSAR, GPS, and dry-tilt data. From 2000 to 2009 the volcano was relatively quiet with geodetic sites slowly moving in toward the SE flank of the volcano, suggesting cooling and contraction of the sill intrusions (Jónsson 2013; Hjaltadóttir et al. 2015). In May 2009 increased seismic activity suggested that a new intrusion was forming under the volcano (Hjaltadóttir et al. 2009; Sigmundsson et al. 2010). A GNSS station on the south flank of the volcano (THEY) moved 10–12 mm toward south during a few weeks period (Fig. 10). After a few months hiatus, the seismic activity and deformation picked up again (Fig. 10), with near-field GNSS stations moving away from the southeast flank of the volcano. This was interpreted as a sill forming at 4 to 6 km depth. Around the end of February 2010, a significant change in horizontal velocities occurred at more distant GNSS stations (Fig. 11, Hreinsdóttir et al. 2012; Thrastarson 2013), with sites moving toward the volcano. The far-field deflation observations indicated that a deep magma reservoir under the volcano was deflating, feeding the ongoing shallow intrusions. In early March, seismic activity

intensified, and rapid deformation and earthquake activity showed upward migration of magma (Hjaltadóttir and Vogfjörð 2010; Tarasewicz et al. 2012). On 20 March 2010, a 300 m-long fissure opened on the east flank of Eyjafjallajökull volcano (Sigmundsson et al. 2010).

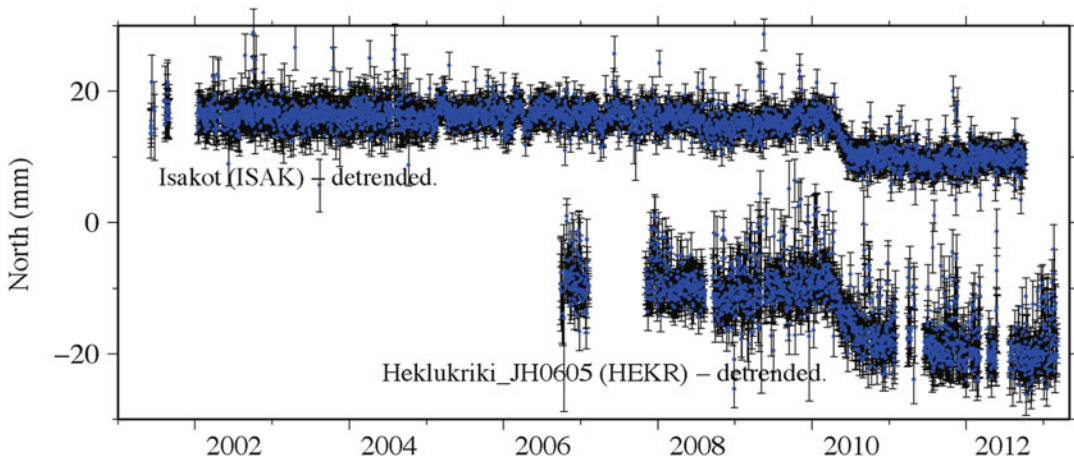
The deformation pattern leading up to the flank eruption was both spatially and temporally variable. During the flank eruption, the deformation almost ceased, and the volcano remained at an inflated state until the eruption ended on 12 April. On 14 April 2010, a more explosive eruption began at the ice-capped summit of the volcano. During this eruption rapid deformation toward the summit and subsidence was observed both with InSAR and at GNSS stations around the volcano (Sigmundsson et al. 2010). The geodetic data are consistent with the deflation of a sill at 4–5 km depth under the summit. The summit eruption ended on 22 May. At around the same time, the far-field sites stopped moving toward the volcano (Hreinsdóttir et al. 2012; Thrastarson 2013).

Following the Eyjafjallajökull eruptions, GNSS stations on the flank of the volcano continued to move into the summit region at a fast decaying rate for a few weeks and months. However, from July 2010 until the end of 2015, the sites closest to the summit moved up and away from the volcano



Volcanoes in Iceland and Crustal Deformation Processes, Fig. 10 The North GPS time series for station THEY at the south flank of Eyjafjallajökull volcano (Hreinsdóttir et al. 2012). A linear term estimated from 2001.0 to 2009.3 has been removed from the time series. A syn-

seismic offset occurred on 17 June 2000 due to a M6.6 earthquake in the SISZ. The yellow region shows periods of intrusive activity under the volcano and red marks syn-eruptive signal



Volcanoes in Iceland and Crustal Deformation Processes, Fig. 11 North component of the GPS time series for Hekla stations ISAK and HEKR (see map Fig. 6). The sites are located about 40 km north of Eyjafjallajökull

volcano. From around 20 February to 23 May 2010, the sites moved south toward Eyjafjallajökull volcano, suggesting pressure drop in a deep magma reservoir (Hreinsdóttir et al. 2012)

(Fig. 11, Jónsson 2013) suggesting recharging of the shallow magma pocket that erupted during the summit eruption.

Uplift and Subsidence

Several intrusion events in Iceland have been detected with geodetic measurements during the past few decades. From 1994 to 1999, inflation of up to 19 mm/yr in the Hengill volcanic system was observed with GPS, leveling, and InSAR measurements. The data were interpreted in terms of a slow inflow of magma into a reservoir at 7 km depth, resulting in intense seismic activity (Sigmundsson et al. 1997b; Feigl et al. 2000). From 2007 to 2008, deformation and seismic activity was detected in the Kverkfjöll volcanic system, at Mt. Upptýppingar. A dike was intruding into the lower crust, north of Vatnajökull ice cap (Geirsson et al. 2010; Hooper et al. 2011). Inflation was also observed at Theistareykir volcano from 2007 to 2008, with maximum uplift of 30 mm/yr, suggesting magma accumulation in a reservoir at 9 km depth (Metzger et al. 2011). In 2009 inflation began in the Krýsuvík volcanic system, suggesting a pressure source at 4–5 km depth. The region inflated for 9 months but then deflated again, reaching pre-inflation state in April of 2010. Another inflation period started in May, with up to 50 mm/yr uplift rate (Michalczevska

et al. 2012). In early 2012 the region started deflating again. Both inflation pulses were associated with increased seismic activity in the crust above the source. Since 2016 Öraefajökull volcano has shown signs of unrest, with increased seismicity and changes in geothermal activity (Jónsdóttir et al. 2018). Subtle signs of inflation have been measured around the volcano, interpreted along with the observed seismic and geothermal activity, as intrusive activity into the volcano (Geirsson et al. 2018).

Geodetic measurements at Krafla volcano show decaying subsidence at the volcano since the Krafla fires, interpreted as cooling and contracting of the magma chamber (Sigmundsson et al. 1997a). Measurements at Askja volcano in north Iceland show rapid deflation since at least 1983 at a slowly decaying rate above a shallow magma chamber (Sturkell and Sigmundsson 2000; Sturkell et al. 2006; de Zeeuw-van Dalfsen et al. 2013). The caldera center subsided almost 1 m from 1983 to 2004. This rapid rate has been interpreted as a pressure drop in the magma chamber, but neither eruption nor known intrusion event took place during this time. GPS measurements show movement into the center of the caldera, and InSAR measurements reveal a few mm/yr subsidence along the fissure swarm (Camitz et al. 1995; Pagli et al. 2006; Pedersen et al. 2009). Askja volcano

had a large explosive eruption in 1875, forming lake Öskjuvatn, a 4.5 km-wide caldera. The eruption was a part of a rifting episode in the volcanic system where an approximately 100 km-long segment of the Askja fissure swarm was activated (Sigurdsson and Sparks 1978; Brandsdóttir 1992). Two smaller episodes took place from 1921–1929 to 1961–1962. Subsidence is also observed at Torfajökull (Scheiber 2008; Geirsson et al. 2012), the largest rhyolitic center in Iceland.

Bibliography

- Albino F, Pinel V, Sigmundsson F (2010) Influence of surface load variations on eruption likelihood: application to two Icelandic subglacial volcanoes. *Geophys J Int* 181:1510–1524. <https://doi.org/10.1111/j.1365-246X.2010.04603.x>
- Allen RM, Nolet G, Morgan WJ, Vogfjörð K, Bergsson BH, Erlendsson P, Foulger GR, Jakobsdóttir S, Julian BR, Pritchard M, Ragnarsson S, Stefánsson R (2002) Imaging the mantle beneath Iceland using integrated seismological techniques. *J Geophys Res* 107:2325. <https://doi.org/10.1029/2001JB000595>
- Altamimi Z, Métivier L, Collilieux X (2012) ITRF2008 plate motion model. *J Geophys Res* 117:B07402. <https://doi.org/10.1029/2011JB008930>
- Árnadóttir T, Sigmundsson F, Delaney PT (1998) Sources of crustal deformation associated with the Krafla, Iceland eruption of September 1984. *Geophys Res Lett* 25:1043–1046
- Árnadóttir T, Lund B, Jiang W, Geirsson H, Björnsson H, Einarsson P, Sigurdsson T (2009) Glacial rebound and plate spreading: results from the first countrywide GPS observations in Iceland. *Geophys J Int* 177:691–716. <https://doi.org/10.1111/j.1365-246X.2008.04059.x>
- Auriac A, Spaans KH, Sigmundsson F, Hooper A, Schmidt P, Lund B (2013) Iceland rising: solid earth response to ice retreat inferred from satellite radar interferometry and viscoelastic modeling. *J Geophys Res* 118:1331–1344
- Auriac A, Sigmundsson F, Hooper A, Spaans KH, Björnsson H, Pálsson F, Pinel V, Feigl KL (2014) InSAR observations and models of crustal deformation due to a glacial surge in Iceland. *Geophys J Int* 198:1329–1341. <https://doi.org/10.1093/gji/ggu205>
- Björnsson A (1985) Dynamics of crustal rifting in NE Iceland. *J Geophys Res* 90:10151–10162
- Björnsson A, Eysteinnsson H (1998) Height changes at Krafla 1974–1995 (in Icelandic with English summary), Report OS-98002/NE9801, ISBN 9979-007-5, Orkustofnun, Reykjavík
- Björnsson H, Pálsson F (2008) Icelandic glaciers. *Jökull* 58:365–386
- Björnsson A, Saemundsson K, Einarsson P, Tryggvason E, Grönvold K (1977) Current rifting episode in North Iceland. *Nature* 266:318–323
- Björnsson A, Johnsen G, Sigurdsson S, Thorbergsson G, Tryggvason E (1979) Rifting of the plate boundary in North Iceland 1975–1978. *J Geophys Res* 84:3029–3038
- Björnsson H, Rott H, Gudmundsson S, Fischer A, Siegel A, Gudmundsson MT (2001) Glacier-volcano interactions deduced by SAR interferometry. *J Glaciol* 47:58–70
- Brandsdóttir B (1992) Historical accounts of earthquakes associated with eruptive activity in the Askja volcanic system. *Jökull* 42:1–12
- Brandsdóttir B, Menke W, Einarsson P, White RS, and Staples, RK (1997) Färoe-Iceland ridge experiment 2. Crustal structure of the Krafla central volcano. *J Geophys* 102:7867–7886
- Buck WR, Einarsson P, Brandsdóttir B (2006) Tectonic stress and magma chamber size as controls on dike propagation: constraints from the 1975–1984 Krafla rifting episode. *J Geophys Res* 111:B12404
- Camitz J, Sigmundsson F, Foulger G, Jahn C-H, Voelksen C, Einarsson P (1995) Plate boundary deformation and continuing deflation of the Askja volcano, North Iceland, determined with GPS, 1987–1993. *Bull Volc* 57:136–145
- Compton K, Bennett RA, Hreinsdóttir S (2015) Climate driven vertical acceleration of Icelandic crust measured by CGPS geodesy. *Geophys Res Lett* 42:743–750. <https://doi.org/10.1002/2014GL062446>
- Compton K, Bennett RA, Hreinsdóttir S, van Dam T, Bordon A, Barletta V, Spada G (2017) Short-term variations of Icelandic ice cap mass inferred from cGPS coordinate time series. *Geochim Geophys Geosyst* 18:2099–2119. <https://doi.org/10.1002/2017GC006831>
- de Zeeuw-van Dalfsen E, Pedersen R, Sigmundsson F, Pagli C (2004) Deep accumulation of magma near crust-mantle boundary at the Krafla volcanic system, Iceland: evidence from satellite radar interferometry 1993–1999. *Geophys Res Lett* 31:L13611
- de Zeeuw-van Dalfsen E, Rymer H, Sturkell E, Pedersen R, Hooper A, Sigmundsson F, Ófeigsson B (2013) Geodetic data shed light on ongoing caldera subsidence at Askja. *Bull Volcanol* 75:1–13. <https://doi.org/10.1007/s00445-013-0709-2>
- DeMets CR, Gordon RG, Argus DF (2010) Geologically current plate motions. *Geophys J Int* 181:1–80. <https://doi.org/10.1111/j.1365-246X.2009.04491.x>
- Drouin V, Sigmundsson F, Hreinsdóttir S, Ófeigsson BG, Sturkell S, Islam T (2014) Crustal movements at a divergent plate boundary: interplay between volcano deformation, geothermal processes, and plate spreading in the northern volcanic zone, Iceland since 2008. *Geophys Res Abstr* 16:EGU2014-10740-3
- Drouin V, Heki K, Sigmundsson F, Hreinsdóttir S, Ófeigsson BG (2016) Constraints on seasonal load variations and regional rigidity from continuous GPS measurements in Iceland, 1997–2014. *Geophys J Int* 205:1843–1858

- Drouin V, Sigmundsson F, Verhagen S, Ófeigsson BG, Spaans K, Hreinsdóttir S (2017) Deformation at Krafla and Bjarnarflag geothermal areas, Northern volcanic zone of Iceland, 1993–2015. *J Volcanol Geotherm Res* 344:92–105
- Dvorak JJ, Dzurlin D (1997) Volcano geodesy: the search for magma reservoirs and the formation of eruptive vents. *Rev Geophys* 35:343–384
- Einarsson P (1978) S-wave shadows in the Krafla caldera in NE-Iceland, evidence for a magma chamber in the crust. *Bulletin Volcanologique* 41:187–195
- Einarsson P (1991) Earthquakes and present-day tectonism in Iceland. *Tectonophysics* 189:261–279
- Einarsson P (2008) Plate boundaries, rifts and transforms in Iceland. *Jökull* 58:35–58
- Einarsson P (2018) Short-term seismic precursors to Icelandic eruptions 1973–2014. *Front Earth Sci* 6:45
- Einarsson P, Brandsdóttir B (1980) Seismological evidence for lateral magma intrusion during the July 1978 deflation of the Krafla volcano in NE-Iceland. *J Geophys* 47:160–165
- Einarsson P, Brandsdóttir B (2000) Earthquakes in the Mýrdalsjökull area, Iceland, 1978–1985: seasonal correlation and connection with volcanoes. *Jökull* 49:59–74
- Einarsson P, Brandsdóttir B, Gudmundsson MT, Björnsson H, Grönvold K, Sigmundsson F (1997) Center of the Iceland hotspot experiences volcanic unrest. *Eos* 78:369–375
- Eliasson J, Larsen G, Gudmundsson MT, Sigmundsson F (2006) Probabilistic model for eruptions and associated flood events in the Katla caldera, Iceland. *Comput Geosci* 10:179–200
- Feigl K, Gasperi J, Sigmundsson F, Rigo A (2000) Crustal deformation near Hengill volcano, Iceland 1993–1998: coupling between volcanism and faulting inferred from elastic modelling of satellite radar interferograms. *J Geophys Res* 105:25655–25670
- Foulger G, Bilham R, Morgan WJ, Einarsson P (1987) The Iceland GPS geodetic field campaign 1986. *Eos* 68:1817–1818
- Foulger GR, Jahn C-H, Seeber G, Einarsson P, Julian BR, Heki K (1992) Post-rifting stress relaxation at the divergent plate boundary in Northeast Iceland. *Nature* 358:488–490
- Fridriksdóttir HJ (2014) Uplift in the Vatnajökull region estimated with GPS geodetic measurements (in Icelandic). B.Sc. Thesis, University of Iceland, Reykjavík, p 26
- Geirsson H, Árnadóttir T, Völksen JW, Sturkell E, Villemin T, Einarsson P, Sigmundsson F, Stefánsson R (2006) Current plate movements across the mid-Atlantic ridge determined from 5 years of continuous GPS measurements in Iceland. *J Geophys Res* 111:B9
- Geirsson H, Árnadóttir T, Hreinsdóttir S, Deciem J, LaFemina PC, Jonsson S, Bennett RA, Metzger S, Holland A, Sturkell E, Villemin T, Völksen C, Sigmundsson F, Einarsson P, Roberts MJ, Sveinbjörnsson H (2010) Overview of results from continuous GPS observations in Iceland from 1995 to 2010. *Jökull* 60:3–22
- Geirsson H, LaFemina P, Árnadóttir T, Sturkell E, Sigmundsson F, Travis M, Schmidt P, Lund B, Hreinsdóttir S, Bennett R (2012) Volcano deformation at active plate boundaries: deep magma accumulation at Hekla volcano and plate boundary deformation in South Iceland. *J Geophys Res* 117:B11409
- Geirsson H, Parks MM, Ófeigsson BG, Drouin V, Li S, Sigmundsson F, Árnadóttir T, Hjartardóttir AR, Lárentinusdóttir M, Einarsson P, Schmidh P (2018) Reawakening of the Öraefajökull volcano in Iceland: deformation signals of stress triggers and intrusive activity. *EGU Gen Assembly* 20:15438
- Gerke K (1974) Crustal movements in the Mývatn- and the Thingvallavatn-area, both horizontal and vertical. In: Kristjánsson L (ed) *Geodynamics of Iceland and the North Atlantic area*. D. Reidel, Dordrecht, pp 263–275
- Gíslason SR, Stefánsdóttir G, Pfeffer MA, Barsotti S, Jóhannsson Þ, Goleczka I, Bali E, Sigmarsson O, Stefánsson A, Keller NS, Sigurdsson Á, Bergsson B, Galle B, Jacobo VC, Arellano S, Aiuppa A, Jónasdóttir EB, Eiríksdóttir ES, Jakobsson S, Gudfinnsson GH, Halldórsson SA, Gunnarsson H, Haddadi B, Jónsdóttir I, Thordarson T, Riishuus M, Högnadóttir T, Dürrig T, Pedersen GBM, Höskuldsson Á, Gudmundsson MT (2015) Environmental pressure from the 2014–15 eruption of Bárðarbunga volcano, Iceland. *Geochem Perspect Lett* 1:84–93
- Grapenthin R, Sigmundsson F, Geirsson H, Árnadóttir T, Pinel V (2006) Icelandic rhythmic: annual modulation of land elevation and plate spreading by snow load. *Geophys Res Lett* 33: L24305, <https://doi.org/10.1029/2006GL028081>
- Grapenthin R, Ófeigsson BG, Sigmundsson F, Sturkell E, Hooper A (2010) Pressure sources versus surface loads: analyzing volcano deformation signal composition with an application to Hekla volcano, Iceland. *Geophys Res Lett* 37:L20310. <https://doi.org/10.1029/2010GL044590>
- Grapenthin R, Siqi L, Ófeigsson B, Sigmundsson F, Vincent D, Hreinsdóttir S, Parks M, Fridriksdóttir H. (2018) Post-eruptive Deformation after the 2014-5 Bárðarbunga dike intrusion and rifting event, Iceland, *EGU General Assembly* 2018 20:15632
- Gudmundsson Ó, Brandsdóttir B, Menke W, Sigvaldason GE (1994) The crustal magma chamber of the Katla volcano in South Iceland revealed by 2-D seismic undershooting. *Geophys J Int* 11:277–296
- Gudmundsson MT, Sigmundsson F, Björnsson H, Högnadóttir T (2004) The 1996 eruption at Gjalp, Vatnajökull ice cap, Iceland: efficiency of heat transfer, ice deformation and subglacial water pressure. *Bull Volcanol* 66:46–65
- Gudmundsson MT, Jónsdóttir K, Hooper A, Holohan EP, Halldórsson SA, Ófeigsson BG, Cesca S, Vogfjörð KS, Sigmundsson F, Högnadóttir T, Einarsson P, Sigmarsson O, Jarosch AH, Jónasson K, Magnússon E, Hreinsdóttir S, Bagnardi M, Parks

- MM, Hjörleifsdóttir V, Pálsson F, Walter TR, Schöpfer MPJ, Heimann S, Reynolds HI, Dumong S, Bali E, Gudfinnsson GH, Dahm T, Roberts MJ, Hensch M, Belart JMC, Spaans K, Jakobsson S, Gudmundsson GB, Fridriksdóttir HM, Drouin V, Dürig T, Adalgeirsdóttir G, Riishuus MS, Pederson GBM, Boeckel T, Oddsson B, Pfeffer MA, Barsotti S, Bergsson B, Donovan A, Burton MR, Aiuppa A (2016) Gradual caldera collapse at Bárðarbunga volcano, Iceland, regulated by lateral magma outflow. *Science* 353(6296):aaf8988
- Hauksson E (1983) Episodic rifting and volcanism at Krafla in North Iceland: growth of large ground fissures along the plate boundary. *J Geophys Res* 88:625–636
- Hjaltadóttir S, Vogfjörð K (2010) Seismic evidence of magma transport in Eyjafjallajökull during 2009–2010, Abstract V21F-02 presented at 2010 fall meeting, AGU, December 13–17, San Francisco
- Hjaltadóttir S, Vogfjörð KS, Slunga R (2009) Seismic Signs of Magma Pathways Through the Crust in the Eyjafjallajökull Volcano, South Iceland. Rep. VI 2009–13. Icelandic Meteorological Office, Reykjavík
- Hjaltadóttir S, Vogfjörð KS, Hreinsdóttir S, Slunga R (2015) Rewakening of a volcano: activity beneath Eyjafjallajökull volcano from 1991 to 2009. *J Volcanol Geotherm Res.* <https://doi.org/10.1016/j.jvolgeores.2015.08.001>
- Hooper A (2008) A multi-temporal InSAR method incorporating both persistent scatterer and small baseline approaches. *Geophys Res Lett* 35:L16302. <https://doi.org/10.1029/2008GL034654>
- Hooper A, Ófeigsson B, Sigmundsson F, Lund B, Einarsson P, Geirsson H, Sturkell E (2011) Increased capture of magma in the crust promoted by ice-cap retreat in Iceland. *Nat Geosci* 4:783–786. <https://doi.org/10.1038/NGEO1269>
- Hreinsdóttir S, Einarsson P, Jónsson S, Sigmundsson F (1998) GPS geodetic measurements around Bárðarbunga volcano, central Iceland, in 1997, Report RH-13-98 Science Institute, University of Iceland, and Report 9802 Nord Volcanol Inst, p 41
- Hreinsdóttir S, Einarsson P, Sigmundsson F (2001) Crustal deformation at the oblique spreading Reykjanes Peninsula, SW Iceland: GPS measurements from 1993–1998. *J Geophys Res* 106:13803–13816
- Hreinsdóttir S, Sigmundsson F, Roberts M, Árnadóttir T, Ófeigsson B, Grapenthin R, Sturkell E, Villemín T, Bennett R, Geirsson H (2012) The 2010 Eyjafjallajökull and 2011 Grímsvötn eruptions: insights from GPS geodesy, geophysical research abstracts. EGU Gen Assembly 14:EGU2012-13577-1
- Hreinsdóttir S, Sigmundsson F, Roberts MJ, Grapenthin R, Arason P, Árnadóttir T, Hólmjárn J, Geirsson H, Bennett RA, Gudmundsson MT, Oddsson B, Ófeigsson BG, Villemín T, Jónsson T, Sturkell E, Höskuldsson Á, Larsen G, Thordarson T, Óladóttir BA (2014) Volcanic plume height correlated with magma pressure change at Grímsvötn volcano, Iceland. *Nat Geosci* 7:214–218. <https://doi.org/10.1038/ngeo2044>
- Jakobsson SP (1972) Chemistry and distribution pattern of recent basaltic rocks in Iceland. *Lithos* 5:365–386
- Jóhannesson H (1982) Overview of the geology of the Snæfellsnes peninsula (in Icelandic). *Arbok Ferðafélags Isl* 151–172
- Jóhannesson H, Saemundsson K (2009a) Geological map of Iceland 1:600000, bedrock geology. Icelandic Institute of Natural History, Reykjavík
- Jóhannesson H, Saemundsson K (2009b) Geological map of Iceland 1:600000, Tectonics. Icelandic Institute of Natural History, Reykjavík
- Jónsdóttir K, Gudmundsson MT, Pfeffer M, Stefánsson A, Parks M, Lecocq T, Heidar Thrastarson R, Geirsson H, Barsotti S, Magnússon E, Pálsson F (2018) Reawakening of Öraefajökull volcano monitored using a multidisciplinary approach. EGU Gen Assembly 20:17969
- Jónsson ÓP (2013) GPS-measurements around the Eyjafjallajökull volcano since the end of the 2010 eruption (in Icelandic with English abstract), BSc. Thesis, University of Iceland
- Juncu D, Árnadóttir T, Hooper A, Gunnarsson G (2017) Anthropogenic and natural ground deformation in the Hengill geothermal area, Iceland. *J Geophys Res* 122:692–709
- Juncu D, Árnadóttir T, Geirsson H, Guðmundsson GB, Lund B, Gunnarsson G, Hooper A, Hreinsdóttir S, Michalczevska K (2018) Injection-induced surface deformation and seismicity at the Hellisheidi geothermal field, Iceland. *J Volcanol Geotherm Res.* <https://doi.org/10.1016/j.jvolgeores.2018.03.019>
- Keiding M, Árnadóttir T, Sturkell E, Geirsson H, Lund B (2008) Strain accumulation along an oblique plate boundary: the Reykjanes peninsula, Southwest Iceland. *Gephys J Int* 172:861–872
- Keiding M, Árnadóttir T, Jónsson S, Decriem J, Hooper A (2010) Plate boundary deformation and man-made subsidence around geothermal fields on the Reykjanes Peninsula, Iceland. *J Volcanol Geotherm Res* 194:139–149
- Kjartansson E, Gronvold K (1983) Location of a magma reservoir beneath Hekla Volcano, Iceland. *Nature* 301:139–141. <https://doi.org/10.1038/301139a0>
- Kristinsson GH (2014) GPS Geodetic Measurements on Snæfellsnes Peninsula (in Icelandic). B.Sc. thesis, University of Iceland, Reykjavík, p 26
- La Femina PC, Dixon TH, Malservisi R, Árnadóttir T, Sturkell E, Sigmundsson F, Einarsson P (2005) Geodetic GPS measurements in South Iceland: strain accumulation and portioning in a propagating ridge system. *J Geophys Res* 110:B11405
- Larsen G, Gudmundsson MT (2014) Volcanic system: bárðarbunga system. Pre-publication extract from the catalogue of Icelandic Volcanoes, published 25 Aug 2014
- Larsen G, Gudmundsson MT, Björnsson H (1998) Eight centuries of periodic volcanism at the center of the Iceland hotspot revealed by glacier tephrostratigraphy. *Geology* 26:943–946

- Larsen GA, Dugmore AJ, Newton AJ (1999) Geochemistry of historical-age silicic tephra in Iceland. *The Holocene* 9:463–471
- Linde AT, Ágústsson K, Sacks IS, Stefánsson R (1993) Mechanism of the 1991 eruption of Hekla from continuous borehole strain monitoring. *Nature* 365:737–740
- Massonnet D, Sigmundsson F (2000) Remote sensing of volcano deformation by radar interferometry from various satellites. *Remote Sens Active Volcanism* 116:207–221
- Massonnet D, Briole P, Arnaud A (1995) Deflation of Mount Etna monitored by spaceborne radar interferometry. *Nature* 375:567–570
- Metzger S, Jónsson S, Geirsson H (2011) Locking depth and slip-rate of the Húsavík Flatey fault, North Iceland, derived from continuous GPS data 2006–2010. *Geophys J Int* 187:564–576
- Michalczevska K, Hreinsdóttir S, Dumont S, Sigmundsson F (2014) Crustal deformation on the Western Reykjanes Peninsula 2009 to 2014 mapped by InSAR and GPS measurements. Institute of Earth Sciences report
- Michalczevska K, Hreinsdóttir S, Árnadóttir T, Hjaltadóttir S, Ágústsdóttir T, Gudmundsson MT, Geirsson H, Sigmundsson F, Gudmundsson G (2012) Inflation and deflation episodes in the Krísvík volcanic system, Abstract, V33A-2843, AGU Fall Meeting
- Moorbath S, Sigurdsson H, Goodwin R (1968) K-Ar ages of the oldest exposed rocks in Iceland. *Earth Planet Sci Lett* 4:197–205
- Niemczyk O (1943) The splitting of Iceland – the German geological, geodetic and geophysical research expedition of 1938 (in German). Konrad Wittwer, Stuttgart, p 180
- Ófeigsson BG, Hooper A, Sigmundsson F, Sturkell E, Grapenthin R (2011a) Deep magma storage at Hekla volcano, Iceland, revealed by InSAR time series analysis. *J Geophys Res* 116:B05401. <https://doi.org/10.1029/2010JB007576>
- Ófeigsson BG, Hreinsdóttir S, Sigmundsson F, Spaans K, Vogfjörð K, Hooper AJ, Sturkell EC, Roberts MJ, Gudmundsson GB, Einarsson P, Árnadóttir T, Jakobsdóttir S, Sigurdsson G, Gudmundsson MT, Geirsson H, Bennett RA (2011b) Increased volcanic unrest at Katla volcano, Iceland? Abstract C53E-2690. AGU Fall Meeting, San Francisco
- Ófeigsson BG, Hreinsdóttir S, Sigmundsson F, Friðriksdóttir H, Parks M, Dumont S, Árnadóttir TH, Geirsson H, Hooper A, Roberts M, Bennett R, Sturkell E, Jónsson S, LaFemina P, Jónsson T, Steinthorsson S, Einarsson P, Drouin V (2015) Deformation derived from GPS geodesy associated with Bárðarbunga 2014 rifting event in Iceland. *Geophys Res Abstr* 17:EGU2015-7691-4
- Óladóttir BA, Larsen G, Sigmarsson O (2011) Holocene volcanic activity at Grímsvötn, Bárðarbunga and Kverkfjöll subglacial centres beneath Vatnajökull, Iceland. *Bull Volcanol* 73:1187–1208
- Óskarsson BV (2009) The Sker ridge on Eyjafjallajökull, South Iceland: morphology and magma-ice interaction in an ice-confined silicic fissure eruption, MSc thesis, University of Iceland
- Óskarsson N, Steinthorsson S, Sigvaldason GE (1985) Iceland geochemical anomaly: origin, volcanotectonics, chemical fractionation and isotope evolution of the crust. *J Geophys Res* 90:10011–10025
- Pagli C, Sigmundsson F, Árnadóttir T, Einarsson P, Sturkell E (2006) Deflation of the Askja volcanic system: constraints on the deformation source from combined inversion of satellite radar interferograms and GPS measurements. *J Volcanol Geotherm Res* 152:97–108
- Pagli C, Sigmundsson F, Pedersen R, Einarsson P, Árnadóttir T, Feigl KL (2007) Crustal deformation associated with the 1996 Gjálp subglacial eruption, Iceland: InSAR studies in affected areas adjacent to the Vatnajökull ice cap. *EPSL* 259:24–33
- Parks M, Heimisson ER, Sigmundsson F, Hooper A, Vogfjörð K, Árnadóttir T, Ófeigsson B, Hreinsdóttir S, Hjartadóttir ÁR, Einarsson P, Gudmundsson MT, Högnadóttir T, Jónsdóttir K, Hensch M, Bagnardi M, Dumont S, Drouin V, Spaans K, Ólafsdóttir R (2017) Evolution of deformation and stress changes during the caldera collapse and dyking at Bárðarbunga, 2014–2015: Implication for triggering of seismicity at nearby Tungnafellsjökull volcano. *Earth Planet Sci Lett* 462:212–223
- Parks M, Sigmundsson F, Sigurðsson Ó, Hooper A, Hreinsdóttir S, Ófeigsson B, Michalczevska K (2018) Deformation due to geothermal exploitation at Reykjanes, Iceland. *J Volcanol Geotherm Res*. <https://doi.org/10.1016/j.jvolgeores.2018.08.016>
- Pedersen R, Sigmundsson F (2004) InSAR based sill model links spatially offset areas of deformation and seismicity for the 1994 unrest episode at Eyjafjallajökull volcano, Iceland. *Geophys Res Lett* 31: L14610. <https://doi.org/10.1029/2004GL020368>
- Pedersen R, Sigmundsson F (2006) Temporal development of the 1999 intrusive episode in the Eyjafjallajökull volcano, Iceland, derived from InSAR images. *Bull Volcanol* 68:377–393
- Pedersen R, Sigmundsson F, Masterlark T (2009) Rheologic controls on the inter-rifting deformation of the Northern Volcanic Zone, Iceland. *EPSL* 281:14–26
- Pinel V, Sigmundsson F, Sturkell E, Geirsson H, Einarsson P, Gudmundsson MT, Högnadóttir T (2007) Discriminating volcano deformation due to magma movement and variable loads: application to Katla subglacial volcano, Iceland. *Geophys J Int* 169:325–338
- Reverso T, Vandemeulebrouck J, Jouanne F, Pinel V, Villemain T, Sturkell E, Bascou P (2014) A two-magma chamber model as a source of deformation at Grímsvötn volcano, Iceland. *J Geophys Res* 119. <https://doi.org/10.1002/2013JB010569>
- Rögnvaldsson ST, Gudmundsson A, Slunga R (1998) Seismotectonic analysis of the Tjörnes Fracture Zone,

- an active transform fault in North Iceland. *J Geophys Res* 103:30117–30129
- Rossi (1997) Morphology of the 1984 open-channel lava flow at Krafla volcano, northern Iceland. *Geomorphology* 20:95–112
- Sæmundsson K (1978) Fissure swarms and central volcanoes of the neovolcanic zones of Iceland. *Geol J Spec Issue* 10:415–432
- Saunders AD, Fitton JG, Kerr AC, Norry MJ, Kent RW (1997) The North Atlantic Igneous Province. In: Mahoney JJ, Coffin MF (eds) Large igneous provinces, continental, oceanic, and planetary flood volcanism, *Geophys. Monogr*, vol 1000. American Geophysical Union, Washington, DC, pp 45–93
- Scheiber SE (2008) Geodetic investigation of Torfajökull volcano Iceland, MSc thesis, University of Witwatersrand Johannesburg
- Segall P (2010) Earthquake and volcano deformation. Princeton University Press, Princeton
- Sigmundsson F (2006) Iceland geodynamics. Praxis Publishing Limited, Chichester
- Sigmarrsson O, Karlsson HR, Larsen G (2000) The 1996 and 1998 subglacial eruptions beneath the Vatnajökull ice sheet in Iceland: contrasting geochemical and geophysical inferences on magma migration. *Bull Volcanol* 61:468–476
- Sigmundsson F, Einarsson P, Bilham R (1992) Magma chamber deflation recorded by the global positioning system: the Hekla 1991 eruption. *Geophys Res Lett* 19:1483–1486
- Sigmundsson F, Einarsson P, Bilham R, Sturkell E (1995) Rift-transform kinematics in South Iceland: deformation from global positioning system measurements, 1986 to 1992. *J Geophys Res* 100:6235–6248
- Sigmundsson F, Vadon H, Massonnet D (1997a) Readjustment of the Krafla spreading segment to crustal rifting measured by satellite radar interferometry. *Geophys Res Lett* 24:1843–1846
- Sigmundsson F, Einarsson P, Rögnvaldsson ST, Foulger GR, Hodgkinson KM, Thorbergsson G (1997b) The 1994–1995 seismicity and deformation at the Hengill triple junction, Ice land: triggering of earthquakes by minor magma injection in a zone of horizontal shear stress. *J Geophys Res* 102:15151–15161
- Sigmundsson F, Hreinsdóttir S, Hooper A, Árnadóttir T, Pedersen R, Roberts MJ, Óskarsson N, Auriac A, Decriem J, Einarsson P, Geirsson H, Hensch M, Ófeigsson B, Sturkell E, Sveinbjörnsson H, Feigl KL (2010) Intrusion triggering of the 2010 Eyjafjallajökull explosive eruption. *Nature* 468:426–430. <https://doi.org/10.1038/nature09558>
- Sigmundsson F, Hreinsdóttir S, Sturkell E, Ófeigsson B, Einarsson P, Roberts M, Grapenthin R, Villemin T, Árnadóttir T, Geirsson H (2014) Grímsvötn volcano, Iceland, 1992–2014: constraints on magma flow in relation to eruptions in 1998, 2004 and 2011. *Geophys Res Abstr* 16:EGU2014-10561
- Sigmundsson F, Hooper A, Hreinsdóttir S, Vogförd KS, Ófeigsson BG, Heimisson ER, Dumont S, Parks M, Spaans K, Gudmundsson GB, Drouin V, Árnadóttir T, Jónsdóttir K, Gudmundsson MT, Högnadóttir T, Fridriksdóttir HM, Hensch M, Einarsson P, Magnússon E, Samsonov S, Brandsdóttir B, White RS, Ágústsdóttir T, Greenfield T, Green RG, Hjartardóttir ÁR, Pedersen R, Bennett RA, Geirsson H, LaFemina P, Björnsson H, Pálsson F, Sturkell E, Bean CJ, Möllhoff M, Braiden AK, Eibl EPS (2015) Segmented lateral dyke growth in a rifting event at Bárðarbunga volcanic system, Iceland. *Nature* 517:191–195
- Sigmundsson F, Einarsson P, Hjartardóttir ÁR, Drouin V, Jónsdóttir K, Árnadóttir T, Geirsson H, Hreinsdóttir S, Li S, Ófeigsson BG (2018) Geodynamics of Iceland and the signatures of plate spreading. *J Volcanol Geotherm Res.* <https://doi.org/10.1016/j.jvolgeores.2018.08.014>
- Sigurdardóttir SS, Gudmundsson MT, Hreinsdóttir S (2015) Mapping of the Eldgjá lava flow on Mýrdalssandur with magnetic surveying. *Jökull* 65:61–71
- Sigurdsson H, Sparks RSJ (1978) Rifting episode in North Iceland in 1874–1875 and the eruptions of Askja and Sveinagjá. *Bull Volcanol* 41:1–19
- Soosalu H, Einarsson P (1997) Seismicity around the Hekla and Torfajökull volcanoes, Iceland, during a volcanically quiet period, 1991–1995. *Bull Volcanol* 59:36–48
- Soosalu H, Einarsson P (2004) Seismic constraints on magma chambers at Hekla and Torfajökull volcanoes, Iceland. *Bull Volcanol* 66:276–286
- Soosalu H, Einarsson P, Þorbjarnardóttir BS (2005) Seismic activity related to the 2000 eruption of the Hekla volcano, Iceland. *Bull Volcanol* 68:21–36
- Spaans K (2011) Deformation at Katla volcano, Iceland, 2003–2009: disentangling surface displacements due to ice load reduction and magma movement using in SAR time series analysis, MSc Thesis, Delft University of Technology
- Spaans K, Hreinsdóttir S, Hooper A, Ófeigsson BG (2015) Crustal movements due to Iceland's shrinking ice caps mimic magma inflow signal at Katla volcano. *Sci Rep* 5:10285
- Sturkell E, Sigmundsson F (2000) Continuous deflation of the Askja caldera Iceland, during the 1983–1998 non-eruptive period. *J Geophys Res* 105:25671–25684
- Sturkell E, Einarsson P, Sigmundsson F, Hreinsdóttir S, Geirsson H (2003a) Deformation of Grímsvötn volcano, Iceland: 1998 eruption and subsequent inflation. *Geophys Res Lett* 30:1182. <https://doi.org/10.1029/2002GL016460>
- Sturkell E, Sigmundsson F, Einarsson P (2003b) 1994 and 1999 unrest and magma movements at Eyjafjallajökull and Katla volcanoes, Iceland. *J Geophys Res* 108:2369
- Sturkell E, Einarsson P, Sigmundsson F, Geirsson H, Ólafsson H, Pedersen R, Zeeuw-van Dalfsen E, Linde AL, Sacks IS, Stefánsson R (2006) Volcano geodesy and magma dynamics in Iceland. *J Volcanol Geotherm Res* 150:14–34

- Sturkell E, Sigmundsson F, Geirsson H, Ólafsson H, Thedorsson T (2008a) Multiple volcano deformation sources in a post-rifting period: 1989–2005 behaviour of Krafla, Iceland constrained by levelling, tilt and GPS observations. *J Volcanol Geotherm Res* 177:405–417
- Sturkell E, Einarsson P, Roberts MJ, Geirsson H, Gudmundsson MT, Sigmundsson F, Pinel V, Gudmundsson GB, Ólafsson H, Stefánsson R (2008b) Seismic and geodetic insights into magma accumulation at Katla subglacial volcano, Iceland: 1999 to 2005. *J Geophys Res* 113:B03212
- Sturkell E, Ágústsson K, Linde AT, Sacks SI, Einarsson P, Sigmundsson F, Geirsson H, Pedersen R, LaFemina PC, Ólafsson H (2013) New insights into volcanic activity from strain and other deformation data for the Hekla 2000 eruption. *J Volcanol Geotherm Res* 256:78–86. <https://doi.org/10.1016/j.jvolgeores.2013.02.001>
- Tarasiewicz J, Brandsdóttir B, White RS, Hensch M, Thorbjarnardóttir B (2012) Using microearthquakes to track repeated magma intrusions beneath the Eyjafjallajökull stratovolcano, Iceland. *J Geophys Res* 117. <https://doi.org/10.1029/2011JB008751>
- Thórarinnsson S (1967) The eruptions of Hekla in historical times. In: Einarsson T, Kjartansson G, Thorarinnsson S (eds) *The eruption of Hekla 1947–1948. I. Soc Sci Islandica*, Reykjavík, pp 1–170
- Thordarson T, Larsen G (2007) Volcanism in Iceland in historical time: volcano types, eruption styles and eruptive history. *J Geodyn* 43:118–152
- Thordarson T, Self S (1993) The Laki (Skaftár Fires) and Grímsvötn eruptions in 1783–1785. *Bull Volcanol* 55:233–263
- Thordarson T, Miller DJ, Larsen G, Self S, Sigurdsson H (2001) New estimates of sulfur degassing and atmospheric mass-loading by the 934AD Eldgjá eruption, Iceland. *J Volcanol Geotherm Res* 108:33–54
- Thrastarson S (2013) Far-field GPS measurements around Eyjafjallajökull in 2010 (in Icelandic with English abstract), BSc thesis, University of Iceland
- Tryggvason E (1968) Measurement of surface deformation in Iceland by precision leveling. *J Geophys Res* 73:7039–7050
- Tryggvason E (1973) Surface deformation and crustal structure in the Mýrdalsjökull area of south Iceland, *J Geophys Res* 78:2488–2497
- Tryggvason E (1982) The NVI magnetoresistor tiltmeter, results of observations 1977–1981. Report, Nordic Volcanological Institute, 8203, p 44
- Tryggvason E (1984) Widening of the Krafla fissure swarm during the 1975–1981 volcano-tectonic episode. *Bull Volcanol* 47:47–69
- Tryggvason E (1986) Multiple magma reservoirs in a rift zone volcano: ground deformation and magma transport during the September 1984 eruption of Krafla, Iceland. *J Volcanol Geotherm Res* 28:1–44
- Tryggvason E (1987) Mývatn lake level observations 1984–1986 and ground deformation during a Krafla eruption. *J Volcanol Geotherm Res* 31:131–138
- Tryggvason E (1989) Ground deformation in Askja, Iceland: its source and possible relation to flow of the mantle plume. *J Volcanol Geotherm Res* 39:61–71
- Tryggvason E (1994a) Surface deformation at the Krafla volcano, North Iceland 1982–1992. *Bull Volcanol* 56:98–107
- Tryggvason E (1994b) Observed ground deformation at Hekla, Iceland prior to and during the eruptions of 1970, 1980–1981, and 1991. *J Volcanol Geotherm Res* 61:281–291
- Tryggvason E (1995) Optical leveling tilt stations in the vicinity of Krafla and the Krafla fissure swarm – Observations 1976 to 1994. Report, Nordic Volcanological Institute 9505, Reykjavík
- Wendt K, Möller D, Ritter D (1985) Geodetic measurements of surface deformations during the present rifting episode in NE Iceland. *J Geophys Res* 90:10163–10172
- Wittmann W, Sigmundsson F, Dumont S, Lavallée Y (2017) Post-emplacement cooling and contraction of lava flows: InSAR observations and a thermal model for lava fields at Hekla volcano, Iceland. *J Geophys Res* 122:946–965
- Wolfé CJ, Bjarnason IT, Van Decar JC, Solomon SC (1997) Seismic structure of the Iceland mantle plume. *Nature* 385:245–247
- Wright TJ, Sigmundsson F, Pagli C, Belachew M, Hamling IJ, Brandsdóttir B, Keir D, Pedersen R, Ayele A, Ebinger C, Einarsson P, Elias L, Calais E (2012) Geophysical constraints on the dynamics of spreading centers from rifting episodes on land. *Nat Geosci* 5:242–250

Volcanic Eruptions, Explosive: Experimental Insights

Stephen J. Lane and Michael R. James
Lancaster Environment Centre, Lancaster
University, Lancaster, UK

Article Outline

Glossary
Definition of the Subject
Introduction
Volcanic Materials
Volcanic Processes
Analogue Approach
Future Directions
Bibliography

Glossary

Conduit Subterranean pathways to the surface created and sustained by the flow of volcanic materials are known as volcanic conduits. In the laboratory, tubes are used as scale equivalents. The two terminologies ('conduits' and 'tubes') should be kept distinct to avoid inadvertent acceptance that a laboratory experiment is a direct replica of a volcanic process.

Magma When rock melts underground it becomes magma. When magma flows on the surface it is known as lava and when it is projected into the atmosphere it forms pyroclasts. Most magma is silicate-based and comprises a three-phase mixture of solid (crystals, called microlites when small), liquid (melt) and gas (bubbles or vesicles).

Rheology When a force is applied to matter it deforms. The rheology of a material describes the relationship between applied force and the material response. Volcanic materials have a wide range of rheological behaviors dependent

on the proportions of gas, liquid and solid, temperature, chemical composition, deformation history and rate of deformation.

Scaling Explosive volcanic eruptions involve temperatures of order 1000 °C, pressures of many MPa and physical sizes from μm to km. Laboratory study of these processes generally involves observation and measurement of processes operating at lower temperature or pressure, or at smaller physical size. Scaling is the means of making laboratory-scale analogues representative of volcano-scale natural processes.

Volatiles Explosive volcanic eruptions are driven by the exsolution of a dissolved gas phase as magma pressure falls. These gaseous species, dominated by water vapor on Earth, are known as volatiles.

Definition of the Subject

The magnitude of volcanic eruptions may be classified using the Volcanic Explosivity Index (VEI) (Newhall and Self 1982; Pyle 2000). The lowest VEI index of zero covers non-explosive activity (such as the production of lava flows) and is generally associated with low magma effusion rates or low viscosity magmas, for example, carbonate-based carbonatites (Woolley and Church 2005). For silicate-based magmas, which are generally more viscous than carbonatites, the boundary between effusive and explosive volcanism is commonly marked by Hawaiian fire-fountaining activity (Parfitt 2004). Magmas with a high silica content, or a lower temperature (and thus higher viscosity), can erupt either explosively or non-explosively as lava flows or dome-building eruptions (Yokoyama 2005). Some eruptions switch between effusive and explosive behavior dependent on prevailing physicochemical conditions (e.g., Ida 2007; Jaupart and Allegre 1991), with magmas tending to become more explosive

at higher effusion rates. High VEI eruptions are, therefore, associated with high flow rates, large erupted volumes and substantial eruption plumes.

Such explosive activity comprises a wide range of phenomena, which result from complex non-linear interactions and feedback mechanisms. Processes that initiate in the subsurface plumbing or conduit system determine the nature of following events and control eruption styles. On emerging from the vent, volcanic material enters the atmosphere and the ensuing interactions are key in determining the consequent transport of volcanic debris that defines the impact on the environment and on human lives and infrastructure. On Earth, our capability to mitigate such volcanic hazard relies in large part on forecasting explosive events, a process which requires a high degree of understanding about the physicochemical factors operating during explosive volcanism.

In common with many science disciplines, the approaches taken to gain an understanding of explosive volcanism have relied on a combination of field observations, theoretical models and laboratory models of materials and mechanisms. Field observations form the ‘ground truth’ for numerical and experimental modeling (and provide vital historical data), but are often very difficult to unambiguously interpret due to their scarcity in both space and time. Numerical modeling complements field observations by providing much greater information density and, hence, the potential to examine processes in great detail. However, the accuracy of the results depends on correctly including all relevant materials properties and system processes in the model and the results can be difficult to verify against often-unobservable field phenomena. Laboratory investigation of the properties of volcanic materials, their modes of interaction and the behaviors of analogous systems provides the third complementary approach; this is the area we focus on here.

Introduction

In the context of explosive volcanic activity, laboratory experiments cover a wide range of research, the most fundamental of which is to

gain an understanding of the properties of the volcanic material involved. Without this to underpin further work, neither accurate computer nor laboratory simulations could be carried out. The materials properties that are relevant to explosive volcanism include the solubility in silicate melts of volatiles (most commonly water and carbon dioxide), melt viscosity as a function of chemical composition, magma rheology as a function of the size, shape and proportion of crystals and bubbles, gas permeability of magma, magma deformation mode (e.g. brittle or ductile) as a function of strain-rate, and volatile diffusivity. Experimental determination of these parameters and their interdependencies results in empirically formulated relationships that can be used in numerical models and compared with the properties of analogue fluids. We first review aspects of the volcanic materials literature in section “[Volcanic Materials](#)”, with the aim of illustrating the nature of molten rock, the complexity of which underpins most explosive volcanic processes.

Experimental modeling of these processes can then build on the materials understanding. Such experiments involve investigation of the behavior of natural volcanic products at laboratory time and length scales, including the response of magma samples to rapid changes in pressure and temperature, the fall behavior of silicate particles in the atmosphere, and the generation and separation of electrostatic charge during explosive eruptions. Here, we encounter the concept of scaling (Barenblatt 2003; Burgisser et al. 2005), which links the applicability of experimental results using small samples of natural materials to natural scales that are often several orders of magnitude larger. For experiments that replace natural volcanic products with analogue materials (for example, in order to relax the experimental pressure and temperature requirements), scaling arguments are crucial. A scaling approach permits adjustment of experimental durations and sizes, whilst maintaining similitude with much larger and longer events. Consequently, analogue experimentation has allowed investigation of explosive phenomena in the broader context of fluid-dynamic behavior, and is a powerful means of giving ‘insight’ into volcanic processes.

Burgisser et al. (2005) review and discuss scaling criteria for experiments and natural phenomena involving dilute multiphase mixtures in magmatic systems, and conclude "... that, despite numerous experimental studies on processes relevant to magmatic systems, some and perhaps most geologically important parameter ranges still need to be addressed at the laboratory scale". This emphasizes that small-scale experiments aiming to mimic larger-scale natural processes need to be considered in two contexts. Firstly, laboratory experiments yield valid results in their own right and provide a means of validating numerical models. Secondly, laboratory experiments are not facsimiles of the natural process, but can give perception into how aspects of natural systems behave provided those aspects are scaled; a process normally considered at design stage. Consequently, comparing experimental results with the natural process must always be undertaken with due caution. The analogue experiment literature is reviewed in sections "[Volcanic Processes](#)" and "[Analogue Approach](#)".

In section "[Definition of the Subject](#)", the combined tactics of field observation, numerical model and experimental model was discussed. In practice, complete overlap of these techniques can be difficult to achieve because the parameters measurable in the field are often incompatible with experimentally accessible variables, which in turn may not coincide with the output from numerical models. Explosive volcanism is also a difficult phenomenon to observe in action, and there are three reasons for this. Firstly, explosions do not happen very often, and the larger the explosion the more infrequent it is, with, for example, repeat timescales of order 10^3 years for VEI 6 events like the Krakatau, 1883 eruption. This makes detailed synevent measurements of large explosive episodes infrequent on the timescale of contemporary volcanology. Secondly, high VEI events tend to destroy near-field observation equipment, making measurements very difficult to obtain even when an event does occur. Thirdly, processes that comprise an explosive event can be difficult to access directly; perhaps a prime example of this is how to obtain data about the nature of flow in the volcanic conduit system during

explosive activity. Direct measurements are not possible and our main source of information is the deformation of the conduit wall created by unsteadiness in the flow, which in turn is calculated from the interpretation of ground motion signals detected by seismometers (e.g., Chouet et al. 2003).

These considerations mean that there are many field observations, numerical and experimental models that await comparison with their companion approaches, however, the detailed study of volcanoes with low VEIs is one area where combined tactics are possible. As an example, in section "[Analogue Approach](#)" we review the contribution that experimental modeling has made to the understanding of fluid flow in the conduit feeding VEI 1–2 explosions at Stromboli Volcano, Italy. Here it has been possible to measure signals from many thousand events, over the period of a few years (e.g., Auger et al. 2006), with a network of seismometers and other instruments, as well as detailed sampling of eruption products (e.g., Lautze and Houghton 2007). Such field data are key to testing both numerical and experimental models of explosive volcanic activity wherever it occurs.

Volcanic Materials

Magma Rheology

Rheology describes the way in which materials deform and flow under the influence of an applied stress. Explosive eruptions only occur because considerable stress is applied to magma. The rheology of natural silicate materials was being measured before World War II (e.g., Birch and Dane 1942), with empirical models (derived from experimental data) becoming established in the 1960s and 1970s (e.g., Bottinga and Weill 1972; Murase and McBirney 1973; Shaw 1972 and references therein).

When stress is applied to a solid it deforms or strains elastically, the strain being the macroscopic result of atoms in the solid moving closer together or further apart. These two parameters are related by a constant called a modulus, for example, in solids under tension $\sigma = E\gamma$, where σ is

tensile stress, γ is tensile strain and E is Young's modulus. Similar relationships exist for deformation under shear or by pressure change. These moduli are likely to be a constant for a range of values of applied stress, but could become dependent on strain, or indeed strain rate.

When stress (σ) is applied to a fluid (the collective term for gas and liquid), viscosity (η) is the parameter that relates the rate of flow (strain rate, dy/dt) to the stress that is being applied to it, giving the relationship $\sigma = \eta dy/dt$. The flow of a fluid is the macroscopic result of fluid molecules moving past each other; therefore, changes that influence molecular motion also affect viscosity. The viscosity can be independent of the rate of flow, a case known as Newtonian behavior, and is best exhibited by gases and liquids of low molecular weight. Fluids whose viscosity is a function of strain-rate are described as non-Newtonian, and many foodstuffs, for instance, exhibit this behavior (e.g. tomato ketchup). Viscosity is also a function of temperature, generally decreasing as temperature increases and molecules move further apart (thermal expansion). An everyday example would be Golden or Maple syrup that becomes

easier to pour on heating. The concept of viscosity fails when molecular interactions are rare, for example in gases at low pressure.

Newtonian Viscosity Natural silicates in the liquid state have been assumed to be Newtonian in behavior and demonstrate a strong viscosity dependence on temperature. The relationship was initially represented by an Arrhenius-type equation, such as $\eta = A \exp[E/RT]$, where the constants A and E are dependent on the chemical composition of the silicate melt (e.g., Murase and McBirney 1973), R is the gas constant and T the absolute temperature. With considerable further laboratory effort in measuring the viscosities of a wide compositional range of natural silicate liquids, it became clear that viscosity was not strictly an Arrhenian function of temperature. Hess and Dingwell (1996) developed an empirical model to describe laboratory values of viscosity for leucogranitic melts as a function of temperature and water content. The compilation of data used by Hess and Dingwell (1996) is shown in Fig. 1, and the empirical equation generated to fit the data is:

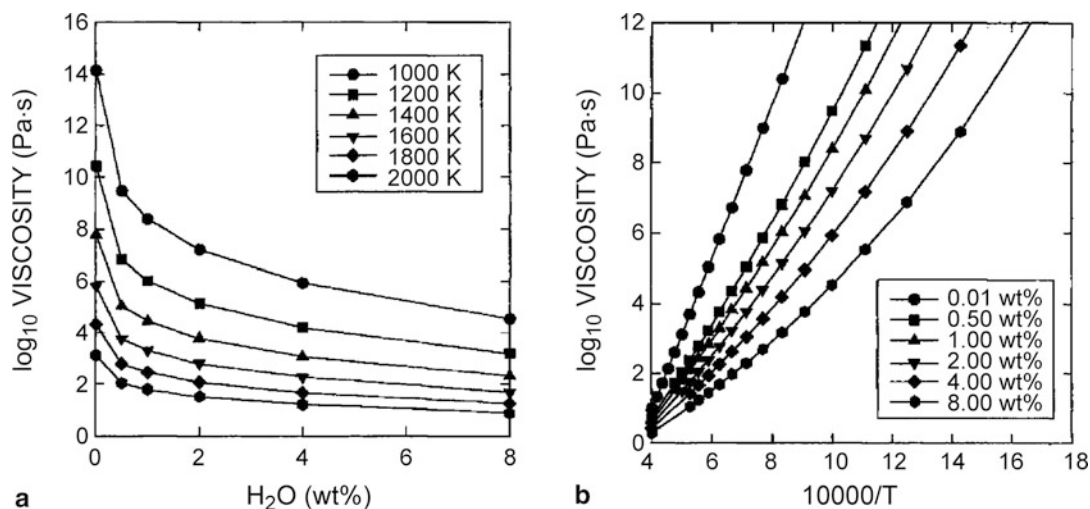
$$\log \eta = \frac{[-3.545 + 0.833 \ln(w)] + [9601 - 2368 \ln(w)]}{\{T - [195.7 + 32.25 \ln(w)]\}},$$

where w is the water % w/w. The constants in this equation will be different for silicate melts of different composition; see Zhang et al. (2003) for an updated and modified version of this model. Water is important here because it is a ubiquitous component of explosive volcanism on Earth. The concentration of water has a large and non-linear impact on viscosity (Fig. 1a) because it acts as a network modifier by breaking Si-Si bonds in a silicate melt, thus decreasing the average molecular weight.

The use of laboratory measurement to provide quantitative understanding of the dependence of viscosity on temperature, water content and other parameters has provided a key component in the numerical modeling of the explosive flow

behavior of natural silicates, as well as pointing the way to identifying fluids for analogue experiments.

Non-Newtonian Viscosity There is laboratory evidence that the viscosity of liquid magmas is not a constant as once assumed, but may be subtly non-Newtonian (Spera et al. 1988). The contents of a volcanic conduit are rarely 100% liquid, with crystals present in many silicate melts below 1250 °C, considerably higher than most eruption temperatures. The effect of crystallization is to significantly change the relationship between viscosity and temperature (Murase 1962), especially as crystal content is both temperature and time dependent. The presence of bubbles is also



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 1 Model curves, based on experimental measurement, of the dependence of viscosity of metaluminous leucogranitic melts on dissolved water content (a) and temperature (b) from a compilation of data (Hess and Dingwell 1996). Note the logarithmic viscosity scale and non-linear nature of the relationships. Explosive

volcanic activity often results in water loss from the melt, and cooling results from expansion and interaction with the atmosphere; melt viscosity increases result. The role this plays on the dynamics of expanding vesiculated flows may be studied by analogue experimental means. (Figure reproduced with kind permission of the Mineralogical Society of America)

ubiquitous in explosive volcanic materials, and these too alter magma behavior in response to an applied stress. The presence of both yields a rheologically complex 3-phase material.

Combined laboratory and field-measurement of the viscosity of lavas from Mount Etna (Pinkerton and Norton 1995) showed that above 1120 °C Newtonian behavior was exhibited. Eruption temperatures ranged from 1084 °C to 1125 °C; therefore, for much of this range the magma behaved in a non-Newtonian fashion as a consequence of the interaction between crystals suspended within the liquid silicate melt. The Etnean basalt was found to be thixotropic, meaning that the viscosity decreased the longer it was deformed, perhaps indicating that crystals were becoming aligned and could consequently move past each other more readily. The basalt also showed pseudoplastic or shear-thinning behavior, meaning that viscosity decreased as the rate of flow increased, perhaps again, because of the dynamics of crystal alignment. Finally, the basalt showed a small but measurable yield stress of <78 Pa, below which it behaved more like a

solid material. The yield stress is likely to be a consequence of crystals weakly binding together into a meshlike structure. The small nature of the yield stress means that the viscosity of these magmas could be described by a power law equation (i.e., the yield stress was ignored) where $\sigma = \eta (\dot{\gamma})^n$. Above 1120 °C, $n = 1$ and Newtonian behavior was observed, but as temperature falls to 1084 °C, n decreases to 0.46, meaning that viscosity decreases as strain rate increases. Pinkerton and Norton (1995) also noticed that physical motion of the silicate melt could induce crystallization as seen by Kouchi et al. (1986). Similar behavior has been found more recently for laboratory measurements of basalts from Japan (Ishibashi and Sato 2007).

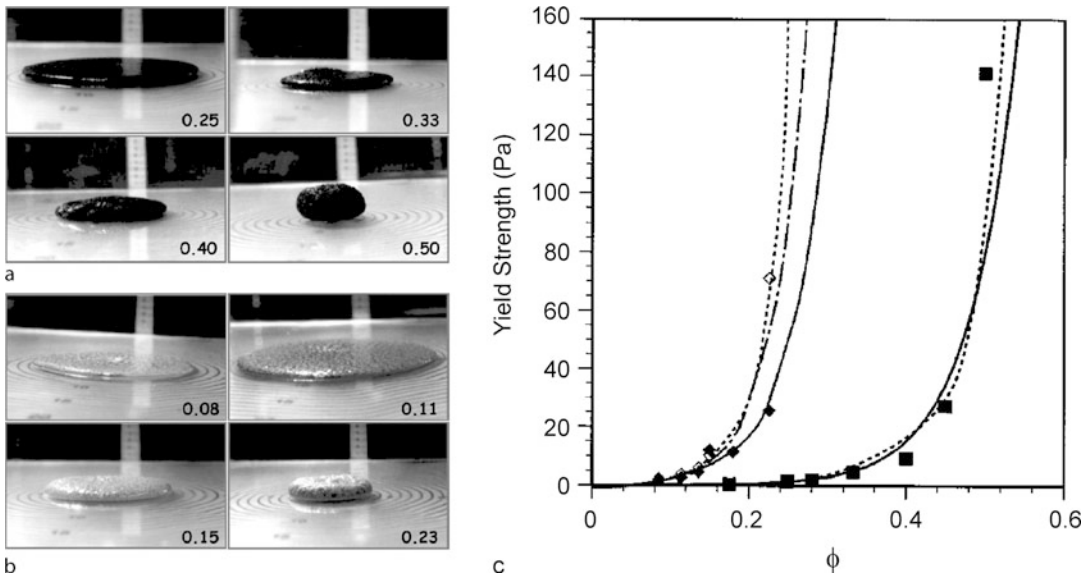
Hoover et al. (2001) showed that crystals induce yield strength in magmas at volume fractions as low as 0.2, although this is strongly dependent on crystal shape. These experiments were carried out on natural basaltic samples, as well as using an analogous system of neutrally buoyant particles in corn syrup (Fig. 2). Natural

samples do, by definition, provide data directly applicable to the natural system, but analogue experiments are complementary because particles have controlled size, shape and concentration. These experiments provide a good example of the kind of complex interplay of time-dependent processes that can occur in magmatic systems that tend only to be quantified by detailed experimentation.

At eruption temperatures, the non-Newtonian behavior of magmas is due in large part to the presence of crystal networks; however, during flow, bubbles must also move past each other and contribute to fluid rheology. Explosive eruptions only occur because bubbles form and expand; therefore, understanding bubble behavior is central to understanding explosive processes. When both bubble size and the fluid flow rate are small, then the surface tension between the silicate liquid and the water vapor in the bubble resists viscous forces and is able to minimize the bubble surface area maintaining spherical bubbles; the bubbles act as if they are solid spheres suspended in the silicate melt. However, if bubbles and flow rate are larger, then surface tension is no longer able to maintain a spherical shape against the viscous drag, and bubbles elongate in the flow field. Under conditions of steady flow, when bubble shape is stable with time, these two behaviors may be distinguished by the capillary number, $Ca = \lambda\gamma'$ where $\lambda = a\eta/\sigma$, a being bubble diameter, η viscosity, γ' strain rate and σ surface tension (e.g., Llewellyn and Manga 2005). When Ca is significantly less than unity, surface tension forces dominate and bubbles remain near spherical. However, when Ca is significantly greater than unity surface tension is overcome by the combination of liquid shear and viscosity, forcing bubbles to elongate. Ca provides an example of a parameter that may be used to scale between volcanic behavior and a laboratory experiment. If bubbles preserved in pumice from an eruption were highly elongated, then any experiment simulating this process would need to also produce highly elongated bubbles and not spherical bubbles. The shape-change undergone by bubbles provides an added complexity over the effects of crystals, whose shapes are generally constant.

Bagdassarov and Dingwell (1992, 1993) experimentally measured the rheological behavior of vesicular rhyolite, using oscillatory rheometry, with a liquid viscosity between 10^9 and 10^{12} Pa s and gas bubble volume fractions in the region between zero and 0.3. Under these conditions, where $Ca \gg 1$, it was shown that vesicular rhyolite decreased in viscosity with increasing gas volume fraction. Such rheological experiments on hot, high-viscosity natural silicate melts are difficult, and the scaling of the small-strain results to larger scales is not straightforward. Using an oscillating viscometer, Llewellyn et al. (2002) measured the viscosity of bubbly Golden Syrup, a fluid analogous to molten silicate with water vapor bubbles, but with liquid viscosity between 10^1 and 10^2 Pa s. The analogous fluid has the advantage of being liquid at room temperature, thereby placing much less stringent demands on any measurement apparatus. Llewellyn et al. (2002) showed the effect of increasing gas volume fraction from zero to 0.5 with small deformations ($Ca \ll 1$) caused viscosity to change by over an order of magnitude. They also discriminated between steady flow (bubbles not changing shape) and unsteady flow (bubbles in the process of changing shape) by proposing the dynamic capillary number $Cd = \lambda\gamma''/\gamma'$, where λ represents the timescale over which a deformed bubble can relax. Under steady flow conditions bubbles tended to increase viscosity, but under conditions of rapidly changing strain rate increasing gas volume fraction reduced viscosity; a result consistent with the oscillatory rheometry results of Bagdassarov and Dingwell (1993). This experimental result largely unified the seemingly contradictory literature that preceded it, and the consequences of these two regimes of behavior are discussed lucidly by Llewellyn and Manga (2005), where the influence of bubbles is shown to change some model outcomes by factors of two or more. This is an excellent example of the role that analogue experimental measurement can play in helping to understand complex and inaccessible natural phenomena.

Viscosity or Modulus? The concept of a relaxation time was encountered above in relation to



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 2 Magma has complex rheology partly because of its crystal content. Hoover et al. (2001) found that basalt with crystals showed solid-like behavior at low stress. Analogue experiments using near-spherical poppy seeds (a) and highly irregular shavings of kaolin/wax (b) in corn syrup demonstrate the role of particle concentration and shape in giving yield strength to solid-liquid mixtures, the number in each panel being the volume fraction of particles. Yield strength increased rapidly (c) for volume

fractions of round poppy seed (filled squares) above about 0.4, allowing extruded material to retain considerable vertical topography (a). The same effect occurs at volume fractions around 0.2 for irregular particles (b) because they interact more strongly with each other than spherical particles. Crystals in magma not only act as nucleation sites for bubbles, but also have a very strong rheological influence in their own right. (Reprinted with kind permission from Elsevier)

how long a deformed (non-spherical) bubble takes to approach the relaxed spherical shape. This same approach can also be applied at the molecular level as liquids flow. During viscous flow, molecules move past each other, and this requires them to be able to reorient or deform. If the timescale of molecular deformation is much shorter than the timescale required by the strain rate during flow, then liquid or viscous behavior ensues. However, if the strain rate is such that molecules cannot reorient or deform, then the material behaves, at least in part, like an elastic solid, with strain being accommodated by bond stretching, or fracture, if yield stress is exceeded.

Materials that demonstrate aspects of liquid and solid behavior are known as viscoelastic. Viscoelasticity was first investigated in the nineteenth century using materials like rubber and glass. Viscoelastic behavior in volcanic materials begins with the onset of non-Newtonian

liquid behavior at deformation timescales about 100 times longer than the relaxation time of silicate molecules in the melt (Dingwell and Webb 1989). The onset of non-Newtonian behavior suggests that there is a range of relaxation timescales for different molecules, and this may be a consequence of silicate melts having a range of molecular weights rather than a single value.

The *timescale* of molecular relaxation is given by the Maxwell relation, *viscosity/modulus*, i.e., the relative weighting of liquid to solid response when stress is applied. In volcanic materials the modulus is effectively constant as temperature and composition change, but viscosity changes by a factor of about 10^9 or more. At low magma viscosities, the relaxation timescale is short and viscous liquid behavior occurs at volcanically plausible strain rates. However, magmas of high viscosity have long relaxation timescales. Here,

the onset of non-Newtonian behavior, and ultimately the glass transition to brittle solid behavior, occurs at much lower strain rates. Dingwell (1996) discusses the relevance of this to explosive volcanic eruptions, and proposes the onset of brittle failure in high-viscosity magma under significant strain rates as a criterion for magma fragmentation. Zhang (1999) showed that magma fragmentation also depends on the number density and size of bubbles within the melt.

Volatile Solubility

Volatiles are defined as those compounds that are dissolved within the matrix of a silicate melt at high pressures, but become supersaturated and, consequently, then exsolve in the supercritical or gaseous state as pressure drops towards atmospheric. On Earth, H₂O is the most common magmatic volatile, with CO₂, SO₂, HCl, and the halogens also present. For water, exsolution occurs into the supercritical state, or gas phase at depths less than about 1 km; therefore, understanding the physicochemical behavior of these compounds at volcanic temperatures and pressures is a key component in the modeling of explosive volcanic eruptions. Water itself has been the subject of extensive laboratory and numerical experiments (with Kalinichev (2001) providing a review) because of its anthropogenic usefulness as well as geological importance. The densities of volatile species are less than that of the silicate melt, for instance at the depths of exsolution the density of water is about 200–400 kg m⁻³, a factor of ~10 less than magma. The bulk modulus of water (order 10⁷ Pa) is much less than that of silicate melt (about 10¹⁰ Pa, Alidibirov et al. 1997). Consequently, as pressure falls, any exsolved volatile phase expands by about three orders of magnitude more than that of the surrounding silicate melt; this is the fundamental driving mechanism of explosive volcanism.

The solubility of volatiles in silicate melts has, and continues to be, the subject of extensive experimentation. Carroll and Holloway (1994) review volatiles in magmas, and Liu et al. (2005) provide a well-referenced introduction to water and carbon dioxide in rhyolite, and give new

experimental data and empirical relationships so important to numerical models of explosive volcanic activity. Empirical H₂O and CO₂ solubility in rhyolitic melt (700–1200 °C and 0–500 MPa) are expressed by Liu et al. (2005) as

$$\begin{aligned} \text{H}_2\text{O}_t &= \frac{(354.94P_w^{0.5} + 9.623P_w - 1.5223P_w^{1.5})}{T} \\ &\quad + 0.0012439P_w^{1.5} \\ &\quad + P_{\text{CO}_2}(-1.084 \times 10^{-4}P_w^{0.5} \\ &\quad - 1.362 \times 10^{-5}P_w), \\ \text{CO}_2 &= \frac{P_{\text{CO}_2}(5668 - 55.99P_w)}{T} \\ &\quad + P_{\text{CO}_2}(0.4133P_w^{0.5} + 2.041 \times 10^{-3}P_w^{1.5}), \end{aligned}$$

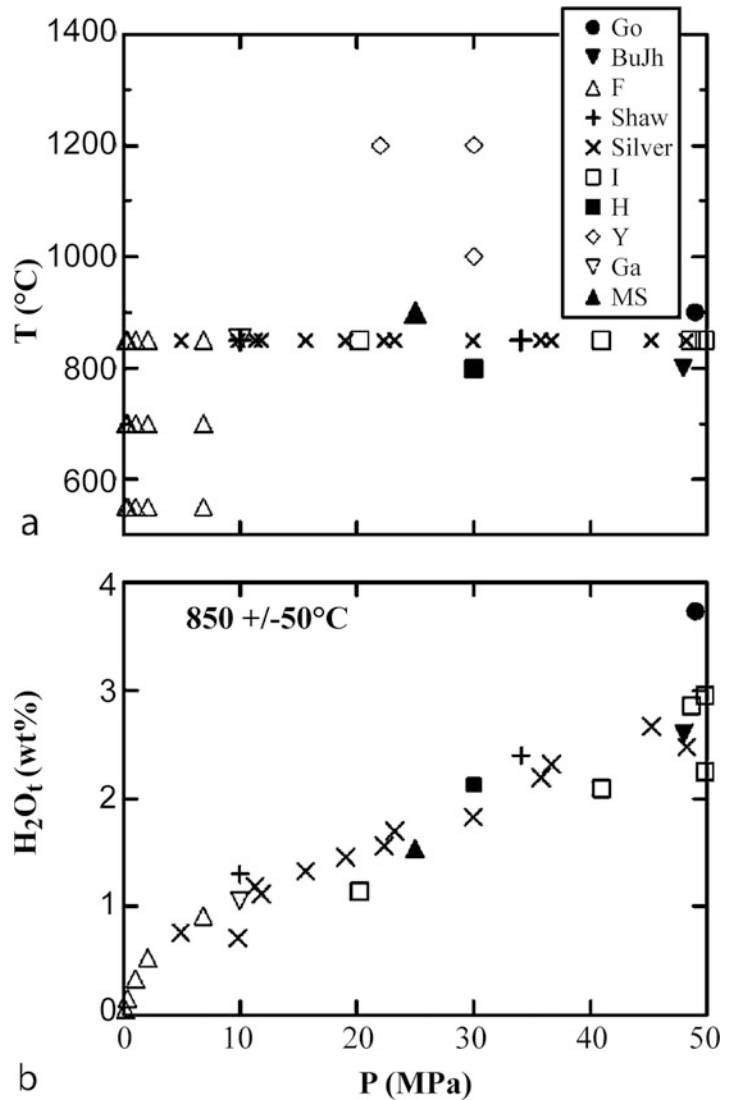
where H₂O_t is the total dissolved H₂O% w/w, CO₂ content is in mass-ppm, *T* is absolute temperature and *P* the partial pressure of water or carbon dioxide.

Figure 3 shows how the solubility of H₂O in rhyolite varies with pressure from a range of experimental work. Experiments focused on explosive volcanism have concentrated on the solubility of water at a temperature of 850 °C, because this is a common eruption temperature for rhyolitic melts. Although water solubility increases as pressure increases, this relationship is not linear, especially at lower pressures, and the nature of the non-linearity is dependent on magma composition (see McMillan 1994 for review). The root of the complexity of water solubility is that water chemically reacts with the silica glass matrix as a network modifier. This reduction in the average molecular weight is responsible for decreasing the viscosity (Fig. 1), as well as having a range of complex geochemical effects (see Behrens and Gaillard 2006 for popular review).

Nucleation and Diffusion

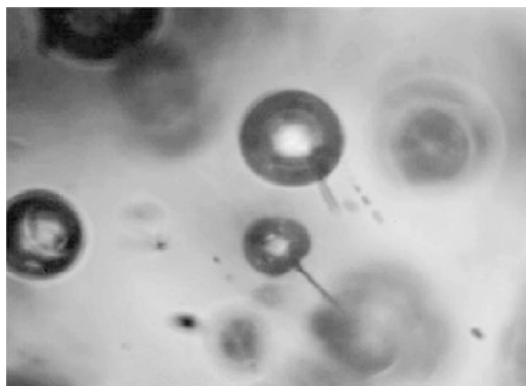
Decreasing the pressure applied to a volatile-saturated magma causes supersaturation of the volatile species. For explosive volcanism to occur, the volatile must make the transition from the dissolved state to a lower-density (and therefore higher volume) exsolved state, i.e., the process of bubble nucleation. Determining the degree of volatile supersaturation required for nucleation

Volcanic Eruptions, Explosive: Experimental Insights, Fig. 3 (a) Shows pressures and temperatures of experiments carried out on rhyolite melts to investigate water solubility as a function of pressure, see Liu et al. (2005) for sources. The bulk of the experiments are carried out at 850 °C because this is a common temperature for rhyolitic eruptions. (b) The solubility of water decreases with pressure, becoming increasingly non-linear below 10 MPa (shallower than about 50 m depth for lithostatic pressure). The transition, and consequent expansion, of water going from the dissolved to the vapor (or supercritical) states as pressure decreases is the driving mechanism of explosive volcanic eruptions. (Reprinted with kind permission from Elsevier)



to take place has been the subject of significant laboratory effort. Magmas are complex chemical compounds that are rarely free of interfaces. Hurwitz and Navon (1994) found experimentally that heterogeneous bubble nucleation dominated, with only small supersaturations of water vapor (<1 MPa, or ~50 m head of magma) being required to nucleate bubbles at the surfaces of Fe-Ti oxide microlites. Other crystal-melt interfaces required supersaturations >10 MPa for nucleation of water bubbles. Even at supersaturations of >130 MPa, nucleation was heterogeneous even in apparently crystal-free melts.

More recently, Gardner (2007) found experimentally that certain crystal interfaces or features acted as nucleation sites, in particular blocky shaped magnetite or the ends of needle-shaped hematite (Fig. 4). Gardner's (2007) study, and others cited therein, indicate that bubble nucleation is a complex function of the composition and temperature of the melt, the composition and morphology of crystals, and the degree of supersaturation prevailing. Accurate modeling of bubble nucleation requires detailed knowledge of the melt as it becomes supersaturated. Nucleation could occur at low supersaturations, with



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 4 As pressure decreases, the melt becomes supersaturated in water. During the decompression of small samples of hydrated rhyolite, bubbles of water vapor nucleated then grew at the ends of needle-shaped hematite crystals, but not along their lengths (scale bar = 20 μm , Gardner (2007)). Such ‘imperfections’ allow heterogeneous bubble nucleation and prevent large supersaturations developing. (Reprinted with kind permission from Elsevier)

conditions never far from equilibrium. Conversely, in the absence of suitable nucleation sites, volatile supersaturations of the order of 100 MPa could develop as pressure in the melt decreases, and when nucleation does occur, rapid bubble growth can be expected.

Once bubbles of volatile have nucleated, growth is promoted by diffusion of volatiles from the supersaturated melt, and by expansion during pressure decrease. Any decrease in pressure also acts to increase volatile supersaturation (Fig. 3), encouraging either new bubble nucleation or, if bubbles are nearby, the diffusion of volatile into existing bubbles (Blower et al. 2001). The diffusivity of volatiles is, therefore, a key parameter in determining the nature of the degassing process and has been reviewed by Watson (1994). Volatile diffusion into a bubble from the local melt surrounding it implies that the bubble will be surrounded by a shell of melt with relatively low volatile content and, thus, with elevated viscosity (Fig. 1a). This viscous melt shell may play an important role in the dynamics of bubble expansion during eruption, further adding to the importance of understanding diffusivity. Zhang and Behrens (2000) experimentally investigated the diffusivity of water in rhyolitic

melts, which, in combination with previous studies, covers the parameter range 400–1200 $^{\circ}\text{C}$, 0.1–810 MPa, and 0.1–7.7% w/w total H_2O content. The resulting empirical relationship between diffusivity of molecular water (D) and pressure (P), temperature (T) and water concentration (X) is given by

$$D_{\text{H}_2\text{O}_m} = \exp \left[\left(14.08 - \frac{13128}{T} - \frac{2.796P}{T} \right) + \left(-27.21 + \frac{36892}{T} + \frac{57.23P}{T} \right) X \right].$$

Such experimental determination of water diffusivity then forms key aspects of models of the behavior of silicate melts that undergo decompression, for instance, in how rapidly bubble pressure recovers after a perturbation (Chouet et al. 2006), or in the nature of the resulting explosive activity (Mason et al. 2006).

Permeability

Erupting magma usually includes gas bubbles, which can approach and interact with each other due to differences in their rise velocities within the melt and due to expansion. If the residence timescale of magma in the plumbing system is long compared to the time required for bubbles to move through it, then the bubbles can be considered to be separating from the melt. This is often the situation in low VEI eruption styles such as Strombolian activity. Here, gas bubbles can ascend through the low viscosity magma relatively rapidly and are able to collide and concentrate, possibly forming bubble rafts or foam layers. Bubble coalescence may then create fewer, but larger bubbles with significantly increased ascent velocity that, on reaching the surface, result in Strombolian activity.

Conversely, if magma residence timescale is short in comparison to bubble rise velocity, then bubble interactions cannot take place so readily. Under these conditions, the rapid depressurization and expansion of the trapped gas bubbles often results in high VEI events, but not universally. The ability of some such magmas to erupt relatively passively indicates that, as they ascend, they are able to lose gas by some other mechanism. One way for this to happen is for the

vesiculated magma to become increasingly able to accommodate gas flow and separation as the volume fraction of bubbles (porosity) increases, allowing gas to move from bubble to bubble to the surface or into the surrounding rock. Such gas escape can be quantified using permeability, which is proportional to the effective pore diameter. Establishing how bubbly magmas become permeable, and the nature of that permeability is, therefore, of key importance to understanding eruption mechanisms.

The permeability of eruption products has only been measured experimentally in recent decades, and by few workers (e.g., Eichelberger et al. 1986; Klug and Cashman 1996; Melnik and Sparks 2002; Saar and Manga 1999). Eichelberger et al. (1986) determined the room-temperature gas permeability of rhyolite with between 37% v/v and 74% v/v porosity. Over this range, permeability increased by over nearly four orders of magnitude from 10^{-16} to 10^{-12} m², precipitating the hypothesis that high permeability can suppress explosive behavior in silicic volcanism, provided gas can escape. Klug and Cashman (1996) made laboratory measurements on 73 pumice samples with porosities ranging from 30% v/v to 92% v/v. Permeability ranged from 10^{-14} to 10^{-12} m², with the relationship that $k \propto \phi^{3.5}$, where k is permeability and ϕ is porosity. Permeabilities at high porosity values agree between these two investigations, but disagreement at low porosities suggests greater complexity. Saar and Manga (1999) demonstrated that there is not a straightforward relationship between permeability and the volume fraction of bubbles. Samples of basaltic scoria, which expand isotropically then cool rapidly, and therefore contain near-spherical bubbles, are well behaved in their permeability-porosity relationship and may be modeled by established percolation theory. However, many multiphase volcanic products do not have spherical bubbles because anisotropic flow causes high values of the capillary number. The elongate nature of the resulting bubbles causes high permeabilities at low porosities (Saar and Manga 1999). However, the permeability, like the porosity, is no longer isotropic, and gas escapes in directions determined by the local direction of flow as well as the prevailing pressure gradient.

Laboratory measurements of volcanic products have been used to scale and validate theoretical models of magma permeability as a function of porosity. These models have, amongst others, been based on analogy with networks of resistors in electric circuits (Blower 2001), and combination of classical approaches with fractal pore-space geometry (Costa 2006). Such models can then be incorporated into numerical simulations of flow in volcanic conduits that test the importance of permeability on explosive-effusive transitions.

Consequences

The consequences of the materials properties of magmas responsible for explosive volcanic flows might now be explored. At pressures higher than the total volatile vapor-pressure, magma remains bubble-free. When pressure is reduced sufficiently below that of the volatile vapor pressure, then bubbles of volatile nucleate in the magma. These bubbles grow, and as bubble density is less than that of the magma, the bulk magma density decreases. This may act to increase magma ascent rate, creating a mechanism of positive feedback. Simultaneously, the reducing volatile content of the magma rapidly increases the liquid viscosity, reducing the strain rate required for viscoelastic and brittle behavior. This process occurs at the length-scale of diffusion into bubbles as well as within the liquid as a whole and, depending on the timescales, crystal growth may also be changing, or indeed driving, the rheological evolution of the system.

These non-linear interactions underpin the complexity of understanding how magma flows, and illustrate the subsurface control on the nature of eruptions. In order to investigate how the properties of volcanic materials interact during the eruption process, experiments of a different nature are required.

Volcanic Processes

Experiments investigating materials properties are mainly designed to measure parameters under static or steady conditions. However, explosive eruptions are not static and only pseudo-steady

at best; therefore, different experiments are needed to explore bulk flow behaviors. Experiments investigating the explosive response of volcanic materials to possible eruption triggers are best conducted using natural silicates that have been well characterized by previous study. These experiments can be designed to minimize the inherent complexity of the overall process by concentrating on specific aspects. Models verified by experiments then contribute to larger models of the whole process.

Explosive Processes

Magma is a hydrated silicate liquid (normally also containing some solid crystals) that, during the explosive eruption process, is converted into silicate fragments ranging in size from 10^1 to 10^{-6} m. The mechanism responsible is fragmentation and understanding fragmentation processes is a major goal of physical volcanology.

Explosive volcanic activity predominantly requires the formation and growth of gas bubbles within magma. Exsolution processes are initiated at pressures generally >10 MPa, and may extend to pressures of several hundred MPa for water. Such pressures result from overburdens of several kilometers and flow experiments carried out at volcanic temperature and pressure (1100–1500 K, up to at least 100 MPa) place severe constraints on containment vessels, measurement transducers and the volume of experimental samples. Another explosive mechanism occurs when volatile species mingle with hot magma at lower pressures resulting in hydrovolcanic activity. This is driven by expansion as liquid turns to gas on heating.

Direct observation of explosive processes is unlikely, but experiments mimicking volcanic conditions give valuable insight. Experimental exploration of explosive processes using natural samples requires that appropriate proportions of gas, liquid and solid are present together with suitable pressures and temperatures.

Hydrovolcanism Experiments investigating the response of hot natural silicate melts to explosive ‘molten fuel-coolant interactions’ (MFCIs) with liquid water have been carried out by Wohletz

(1983) and Zimanowski et al. (1986). MFCIs are not confined to geological processes and are a dangerous industrial hazard, for example, on 4 November 1975, at the Queen Victoria Blast Furnace, Appleby–Frodingham Steelworks, Scunthorpe, UK, up to 90 tons of molten metal were thrown over a wide area after an MFCI explosion involving about 2 tons of water and 180 tons of molten metal. Berthoud (2000) reviews MFCIs from the perspective of the nuclear power and other industries, but the physics described also applies to hydrovolcanism.

Volcano-specific research into MFCIs has studied the size and morphology of silicate particles from volcanic eruptions and compared them to those generated experimentally by MFCI processes (Wohletz 1983). The ratio of water to magma was found to be a key parameter in determining the nature of the particles resulting from the explosive process, with particle size reducing as explosivity increases. Such experiments provide insight into hydrovolcanic eruptions where magma may interact explosively with surface water (Surtseyan eruptions) or ground water (phreatomagmatic eruptions). Two processes dominantly produce the explosivity: (1) the rapid ‘flashing’ of liquid water to vapor; and (2) rapid cooling of the silicate melt causing thermal shock. Zimanowski et al. (1997) studied how explosive fragmentation of a remelted volcanic rock could be interpreted from the size and shape of expelled silicate fragments. Two methods of fragmentation were used: (1) injection of air at 5 or 10 MPa into the base of the crucible containing the melt (at 1653 K); and (2) initiation of a MFCI using liquid water and a triggering perturbation. With an air injection at 5 MPa pressure, fine particles were uncommon and the particles generally rounded; however, the particle size distribution suggested that two unspecified fragmentation processes had operated. Three unspecified fragmentation modes were implied for the 10 MPa air injections, with finer, but still rounded, particles produced. Interaction with water (MFCI) also had three fragmentation mechanisms, but produced the finest particles, many of which were blocky. The MFCI experiments resulted in a much more complex fragmentation process than the air injection

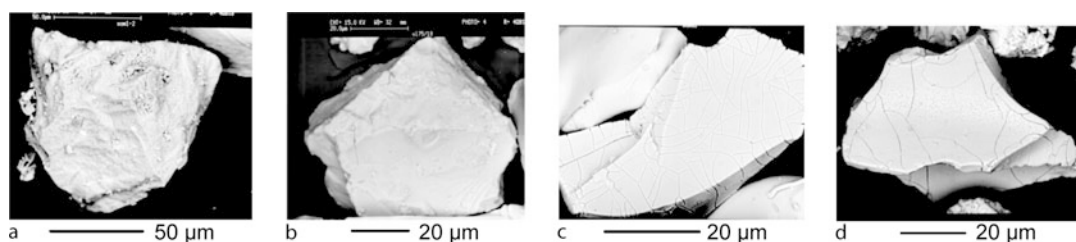
experiments (Zimanowski et al. 1997), with a key difference being the rapid cooling of the silicate melt by interaction with water. The combination of rapid cooling and high strain rates generated by large pressures is considered to cause brittle failure of the melt and generate the fine, blocky fragments during the MFCI experiment.

Although not hydrovolcanism, the air-injection method of Zimanowski et al. (1997) has been used by Taddeucci et al. (2007) to experimentally investigate the role of crystal volume fraction during the explosive fragmentation of magma. This is of interest because ascending magmas are thought to have both axial and radial heterogeneity in their crystal contents, both before and during explosive activity. Preliminary results showed a strong correlation between the fragment morphology and the crystal content in both experimental and natural samples, with increasing crystallinity promoting brittle fracture. This suggests a number of interpretations: (1) a higher viscosity liquid phase naturally exists in more crystalline samples; and/or (2) strain-rates in the smaller liquid volume of high crystallinity samples are higher, i.e., there is less liquid to accommodate a similar motion; and/or (3) the macroscopic rheological behavior of the two-phase mixture promotes brittle processes (Hoover et al. 2001).

Comparisons between experimental and natural particles generated by MFCI experiments (Büttner

et al. 1999) show that quite subtle surface features can be used to distinguish between events involving either partial or complete water vaporization (Fig. 5). Where vaporization is complete, both experimental and natural particles are blocky and equant in shape, and a brittle fragmentation process is implied from observations of stepped surfaces on the particles, created by mechanical etching. Particles resulting from incomplete vaporization of water show similar features, except that their surfaces host a network of cracks that are considered to be the result of rapid surface cooling (thermal shock), created by interaction with excess liquid water. Büttner et al. (1999) comment on the experimental relevance of their research; “In general, the artificial pyroclasts produced in the experiments cover all grain sizes and shape properties of the natural ones, thus illustrating the quality of the experimental scaling in terms of geometry and energy release.” Such phenomenological similarities are a powerful indicator that experimental processes are mimicking the intended natural processes, and this is additional evidence of similarity. However, the experiments did not mimic all the post-explosion fragmentation processes that lead to the final natural product, demonstrating that care must be taken in comparing the outcomes of controlled experiments with the final product of the volcanic process.

Grunewald et al. (2007) introduced a chemical aspect to MFCI experiments by using near-



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 5 Magma and water may interact explosively during phreatomagmatic eruptions. Molten fuel-coolant interactions (MFCI) or magma-water interactions (MWI) can be carried out experimentally (Fig. 6). (a) shows a fragment generated by experimental MWI (Büttner et al. 1999) where all water was vaporized. (b) shows a fragment from a natural phreatomagmatic eruption that has not interacted with excess water. (c) shows an

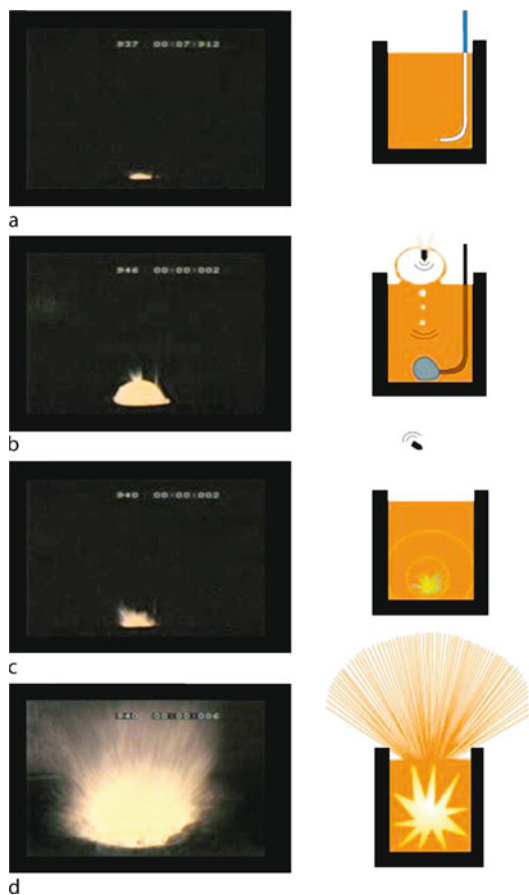
experimental fragment that ‘erupted’ through excess water, and (d) the product of phreatomagmatic activity thought to have interacted with excess water. The phenomenological similarities between (a) and (b), and (c) and (d) provide powerful evidence that experiments are mimicking the processes that occurred during natural phreatomagmatic eruptions. (Reprinted by kind permission from Macmillan Publishers Ltd: Nature (Büttner et al. 1999), copyright (1999))

saturated NaCl solution as well as pure water, with the intention that the addition of other species would add complexity. Experimental methods were similar to that of Zimanowski et al. (1997) (Fig. 6), with the triggering of a rapid thermal interaction by the injection of hot molten silicate with cool liquid brine. It was found that NaOH and HCl were formed during the heating of the NaCl solution, and that explosion energies were

lower than with pure water. The lower explosion energies resulted in more rounded and larger silicate fragments as found by Zimanowski et al. (1997). The reason for the reduced physical energy of fragmentation was attributed to energy coupling into the endothermic chemical process that produced NaOH and HCl, but the heating of brine may well also have introduced a liquid NaCl phase (1074–1738 K) that altered the physical MFCI process. These experiments raise the interesting prospect of being able to use the chemistry of volcanic gases and particle surfaces to make inferences about physical processes occurring during hydrovolcanism.

A different experimental approach to studying magma-water interactions (MWI) was used by Trigila et al. (2007) in which experiments could be carried out at pressures up to 200 MPa, and temperatures up to 1473 K. In contrast to the method of Zimanowski et al. (1997), where water is injected into hot (1650 K), liquid-dominated silicate melt at 0.1 MPa, Trigila et al. (2007) ‘infused’ water into permeable sintered samples of solid-dominated basalt at lower temperature (1100 K), but high pressure (8 MPa). The samples were manufactured by grinding basalt to grain sizes of $<10\ \mu\text{m}$ (powder) or about 0.7 mm (granular). These were heated to experimental temperatures and allowed to sinter for 30 min to form a permeable cylinder (about 2 cm in diameter and 6 cm in length) into which water was introduced. Only the more permeable granulated samples underwent spontaneous explosive behavior on mixing with water, generating particles similar in size to the original grains. These exploratory experiments indicate the importance of permeability to MFCI processes where groundwater could interact with hot, porous country rock as well as magma. We have already explored the importance of permeability in allowing exsolving volatiles to escape and reduce explosivity; under different circumstances we now we see that permeability can also increase explosivity.

Exsolution The nature of geological materials at pressures and temperatures existing within the crust and mantle has been the subject of considerable research by petrologists. Experimental



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 6 Schematic diagrams and frames from high-speed video illustrate one experimental method of simulating explosive phreatomagmatic eruptions (Grunewald et al. 2007). (a) Pre-experiment conditions with water injection tube in crucible of hot (1300 °C) basalt melt. (b) 15 ml of water is injected, with the air expelled from the tube forming a bubble. The bubble was impacted by an air gun pellet to trigger (c) the explosive mixing of water and melt. (d) Rapid transition of water from liquid to vapor results in pressure rise. The melt is fragmented by the subsequent rapid expansion. (Reprinted with kind permission from Elsevier)

petrology generally studies samples of material in small (millimeter-scale) capsules that are sealed, then pressurized up to tens of GPa and heated up to 2000 K; easily covering volcanically relevant conditions. Adaptation of this experimental technology to volcanic processes by Hurwitz and Navon (1994) was used to study the response of hydrated rhyolite to decompressions up to 135 MPa (equivalent to about 5 km silicate overburden) at eruption temperatures (780–850 °C). Extended flow processes cannot be simulated with such small samples at these experimentally challenging conditions, but the pseudo-static response of hydrated silicate melt to pressure decreases of varying rate and magnitude provides important dynamic information about the processes of volatile exsolution (bubble nucleation and growth) and diffusion, as well as crystal growth. Hurwitz and Navon (1994) revealed that nucleation of water bubbles (as a supercritical fluid or gas) in natural magmas is likely to be dominated by various heterogeneous processes operating over a range of supersaturations.

Lyakhovsky et al. (1996) carried out further analysis of the experimental results of Hurwitz and Navon (1994) to reveal the controls on bubble growth following nucleation. Numerical modeling of the experimental data showed that diffusion of water from the melt into the bubbles controlled bubble growth at modest supersaturations in hydrated melts (3–6% w/w water, viscosity 10^4 – 10^6 Pa s). The viscosity of the melt played a controlling role only when bubble surface area was small in the initial stages (<1 s) of growth. At high supersaturations, the rapidly increasing area of bubble wall physically moves larger volumes of melt closer to the bubble and volatile advection is a significant contributor to bubble growth under certain conditions. As bubbles grow, the water content of the melt declines and the melt viscosity rapidly increases (Fig. 1a). Bubble growth then again becomes influenced by viscosity. These experimental results indicated that at water concentrations >3% w/w in rhyolite, equilibrium degassing was a good approximation. However, as water content declines then equilibrium is not maintained and melt some distance away from a bubble wall, and therefore out of

range of significant diffusion, becomes supersaturated in water. Lyakhovsky et al. (1996) postulated that these supersaturated melt pockets then encourage additional episodes of bubble nucleation; a process supported by Blower et al. (2001) using analogue experimentation and computer modeling. Constrained rapid-decompression experiments provided key insights into the role of continuous nucleation in near-equilibrium degassing of high-viscosity, low-diffusivity melts, with supersaturation limited to that required by heterogeneous nucleation sites.

The role of high viscosities (10^7 – 10^9 Pa s) on bubble nucleation was studied by Gardner (2007), using methods similar to those of Hurwitz and Navon (1994). Heterogeneous and continuous nucleation of water bubbles in a rhyolite with less than 1% v/v microlites of Fe–Ti oxides and plagioclase were again exhibited (Fig. 4). Increasing viscosity and decreasing supersaturation were found to increase bubble nucleation times from a few seconds to greater than 100 s. The controlling parameter for nucleation rate was found not to be temperature, diffusivity or viscosity, but the surface tensions of interfaces between melt, crystal and bubble.

All these experiments were fully sealed, with no opportunity for the exsolving volatile phase to escape. This condition was relaxed by Burgisser and Gardner (2005) who decompressed 3 mm diameter, 7 mm long cylinders of rhyolite hydrated within gold capsules at 150 MPa and 825 °C, with some capsules modified to incorporate a sink for escaping water (pseudo-open conditions). Three bubble growth regimes were identified that depended on balances between diffusivity and viscosity in relation to the rate of sample decompression. The open system experiments showed that, given sufficient time (about 180 s here) water bubbles begin to coalesce once bubble volume fraction exceeds about 43%; note that the permeability of natural samples tends to decline rapidly below about 40% porosity. This is important because bubble coalescence promotes an increase in permeability as preferential degassing pathways form in the melt. The limited

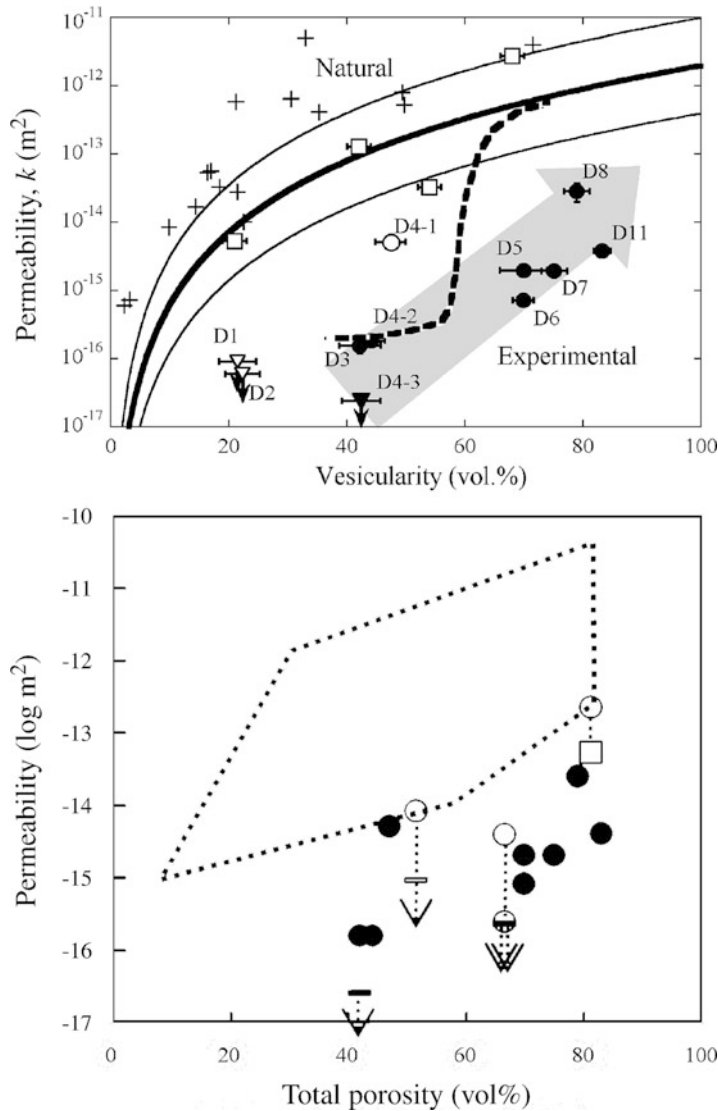
volume of the sample, and its confining capsule, prevents the establishment of the range of flow behaviors that exist in volcanic conduits during explosive eruptions. However, the experiments of Burgisser and Gardner (2005) provide useful insight into the role of flow on bubble coalescence for this very reason; features seen in the experimental samples are likely to have been difficult to distinguish in any experiments aimed at flow behavior. This demonstrates the usefulness of experiments for investigating spatially and temporally small (but tractable) aspects of a complex bigger picture. The results of such experiments can be used to verify models of the partial-process, with the facility to eventually combine these into a description of the full process.

The link between porosity and permeability was further explored by Takeuchi et al. (2005) who isothermally decompressed 5 mm diameter, 15 mm long samples of Usu dacite in gold tubes from 150 MPa at 900 °C, to pressures ranging from 50 to 10 MPa. Samples were then measured for permeability along the tube axis and, with their bubble fraction ranging from 22 to 83% v/v, covered the porosity range over which permeability has been measured in natural samples. However, the experimental samples showed significantly lower permeabilities than natural samples, except for the low porosity data of Eichelberger et al. (1986), and for experiments where bubbles were elongated along the permeability measurement direction (Fig. 7). This suggests that flow processes act to increase permeability by 1–2 orders of magnitude, and that most of the natural samples measured have undergone significant flow; or as Takeuchi et al. (2005) state “The quenched experimental products can be considered as snapshots in vesiculating magma during decompression without deformation of vesicular structure in the late stages of eruption”. Describing experiments as snapshots of a larger, more-complex process is an apt analogy. More recently, Takeuchi et al. (2008) have revisited their experimental samples and, after removing the gold tubes, found that the sample permeabilities were further reduced (Fig. 7), especially at low vesicularity. The

original measurements had included some gas leakage between the silicate sample and the gold tube and, when operating at the limits of permeability detection, the leakage added a significant systematic error to the measurements. The new results further emphasize the importance of the flow history of vesicular materials in the development of permeability, and illustrate the complexity of interactions in the magmatic degassing process that determine the eruptive behavior.

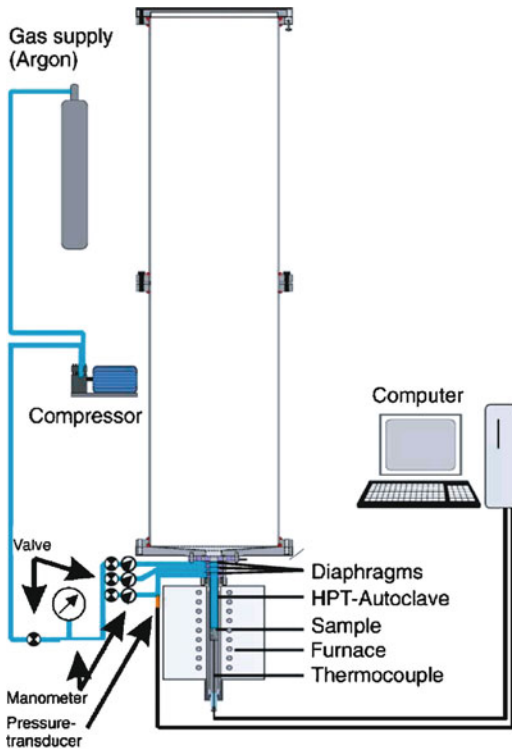
Decompression-Driven Flows At Mt. St. Helens in 1980, a landslide acted to reduce the pressure applied to a body of vesiculated magma. Because the vesicles contained gas, the reduced pressure drove extensive magma expansion. This created the explosive lateral blast that preceded sustained explosive activity on May 18th 1980. Similar behavior may also be exhibited when portions of lava domes collapse. The response of vesiculated magma samples to relatively rapid decompression (10^{-3} – 10^1 GPa s⁻¹ (Alidibirov and Dingwell 2000)) forms a major area of experimental volcanology.

Laboratory apparatus known as a ‘fragmentation bomb’ (Fig. 8) was conceived by Alidibirov and Dingwell (1996a, b) and has been developed by others since. This shock-tube apparatus rapidly decompresses experimental samples at 850 °C from up to 35 MPa to 0.1 MPa (Spieler et al. 2004). The samples, cylinders of approximate dimensions of 2 cm diameter by 20 cm length, are natural volcanic rocks heated to the required temperature in an atmosphere of Ar within a pressure vessel. On reaching the temperature of interest, the sample is further pressurized with Argon until a system of calibrated rupture discs fail. On failure, the sample rapidly decompresses (1 – 100 GPa s⁻¹) and any explosive products freely expand into the collecting tank for further analysis. The use of Ar for these experiments removes any additional complexities that may arise from the nucleation, diffusion and growth of water bubbles allowing the detailed study of the mechanical response of vesiculated silicates to a rapid pressure drop.



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 7 As bubble volume fraction (vesicularity) increases, more bubbles coalesce to allow easy gas exchange between them. The result of this is increasing sample permeability with increasing vesicularity. Permeability allows gas to escape from the magma and reduces explosive potential, making permeability an important parameter in understanding the transition from effusive to explosive volcanism. Experiments carried out by Takeuchi et al. (2005) (circles and squares with error bars, the solid and open symbols indicate isotropic and deformed vesicular textures of the products, respectively) are compared with measurement of natural samples in the top panel (dashed curve indicates the relationship of Eichelberger et al. (1986), the bold solid curve with two fine satellite lines indicates the

relationship of Klug and Cashman (1996) with the upper and lower limit of the scattered data, the crosses and squares represent permeabilities of Melnik and Sparks (2002), and Takeuchi et al. (2005) respectively). Permeabilities have been corrected at a later date because of measurement artefact (bottom panel, Takeuchi et al. 2008). The trend of increasing permeability with vesicularity is demonstrated, but experiments generally show lower permeability than natural samples. One explanation for this could be that most of the natural samples have been subject to significant flow, whereas the experimental samples were generally not (open symbols show some evidence of flow). The role that flow plays in the development of permeability remains unclear. (Reprinted with kind permission of AGU)



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 8 Schematic diagram and photograph of ‘fragmentation bomb’ shock-tube apparatus (Kueppers et al. 2006). Natural samples are pressurized up to 50 MPa with gas and heated up to 850 °C in the autoclave. Samples can be used at room temperature with a transparent experimental section for high-speed imaging (Spieler

et al. [III] (Fig. 9); Taddeucci et al. 2006 (Fig. 22)). The diaphragms are ruptured to decompress the sample to atmosphere (0.1 MPa) and the resulting flow is collected in the large tank. Pressures, images and fragment morphology then provide information on flow dynamics. (Reprinted with kind permission from Elsevier)

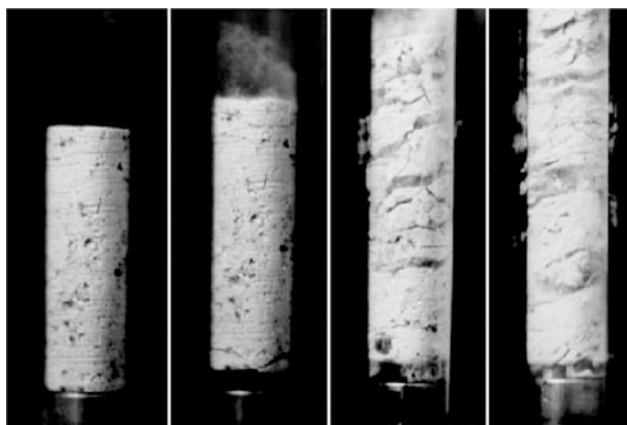
Alidibirov and Dingwell (2000) provide a review of the findings of experiments to that date. Samples of Mt. St. Helens Dacite from the May 18th 1980 eruption were decompressed by up to 18.5 MPa at temperatures up to 950 °C. Observation of the resulting fragments showed angular material characteristic of brittle failure of the samples. Calculations of the strain rates imposed on the samples indicated that, even at the highest temperatures used, the dacite would behave as a brittle solid. Alidibirov and Dingwell (2000) analyze three mechanisms that could explain explosive fragmentation of the sample. The dacite samples had interconnected pores and were, therefore, permeable. Rapid removal of the Ar gas above the sample creates a pressure gradient within the sample. If permeability is very high

then the Ar gas can readily escape and pressure gradients will be relatively low. At the other extreme, if bubbles are not interconnected then pressure gradients within the sample will be maximized (note, the experimental samples must have some permeability in order to pressurize them with Ar, but this does not need to be the case volcanically). In reality, the samples have a degree of permeability that allows pressure gradients to develop within the dacite. The pressure gradient was found to be at its highest at the fragmentation front, where the yield strength of the sample was exceeded. The fragmentation front was found to intrude the sample at tens of m s^{-1} , much slower than the sound speed and known as the fragmentation wave. In this instance, fragmentation can be described as a pressure-gradient driven change in

flow regime from permeable filtration flow where only the gas moves, to two-phase flow where gas and solid are both in motion.

Martel et al. (2000) synthesised 2 cm long by 2 cm diameter samples of hydrated haplogranitic glass with water contents ranging from 1.4 to 5.7% w/w. Water bubbles were nucleated and grown within the samples under a range of temperatures and pressures within the fragmentation bomb apparatus (Fig. 8) on timescales of about 1 h, giving vesicularities ranging from zero to 0.91 v/v. Samples that developed large bubbles and high vesicularities underwent significant expansion and tended to develop elongated, or tube, bubbles during this process. Subsequent rapid depressurizations of between 5 and 18 MPa resulted in explosive fragmentation of the pre-foamed samples purely from expansion of the gas phase within the bubbles; timescales were too short for any significant further diffusion and exsolution of water from the glass. The resulting fragments were collected and analyzed

for size, shape and vesicularity. The primary fragmentation process was found to be a sequential brittle spalling of the upper sample surface exposed to the decompression and was imaged by Spieler et al. (2003) using Santorini pumice at room temperature in a transparent pressure vessel (reproduced here in Fig. 9). Low vesicularities and small decompressions were found to yield larger fragments than high vesicularities and large decompressions, a result consistent with the amount of energy released during the decompression process. The fragment size distributions may be dominated by the primary fragmentation mechanism (surface spalling), but impacts between fragments, and with the apparatus wall, may increase the proportion of the smallest fragments at the expense of the largest, especially with large decompressions (giving high fragment velocities). Interestingly, the presence of tube bubbles acted to decrease fragment size, demonstrating again that flow history is important in determining the products of explosive events.



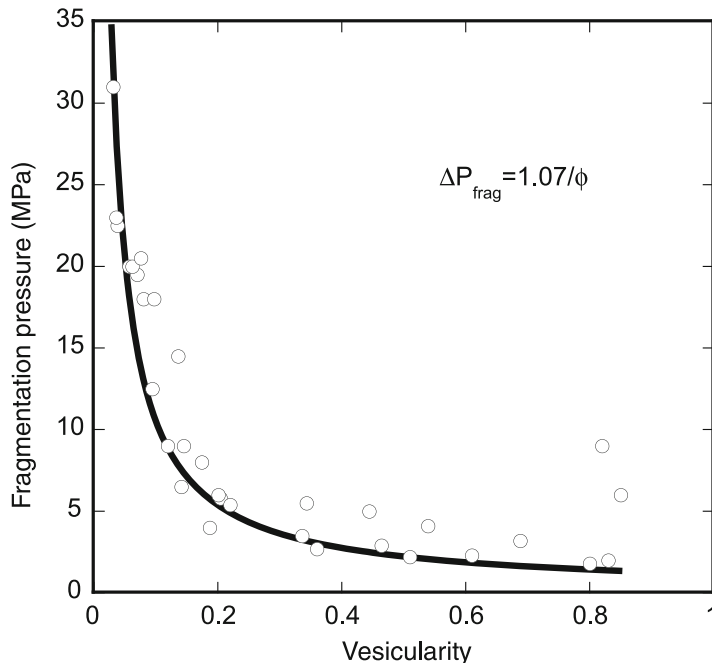
Volcanic Eruptions, Explosive: Experimental Insights, Fig. 9 The rapid decompression of a sample of Santorini pumice, at room temperature, by at least 9 MPa (three times higher than at 850 °C) in the ‘fragmentation bomb’ (Fig. 8) results in spalling fragmentation (Spieler et al. 2003). The first frame shows the 50-mm length by 20-mm diameter sample just prior to decompression. 0.25 ms after diaphragm rupture (second frame) the ejection of fine dust from the sample surface indicates that pressure is falling in the upper reaches of the sample and gas is starting to escape from the pores. The combined elastic response of the sample and apparatus has detached the sample from the holder, fracturing the sample base.

0.5 ms after decompression (third frame), spalling fragmentation propagates through the sample perpendicular to the axis of decompression. Fragmentation results from the build up of a pressure gradient within gas in the upper sample sufficient to exceed the tensile strength of the bubbly solid. After 0.75 ms (fourth frame) primary fragmentation is complete and fragments are ejected into the large tank (Fig. 8). The timescale of this transient flow is of order 1 ms, but such visualizations give great insight into possible volcanological fragmentation mechanisms. (Reproduced with kind permission from Springer Science and Business Media)

In a refinement of these experiments, Martel et al. (2001) investigated the effect of adding relatively inert alumina particles to the hydrated haplogranitic samples as an analogue to crystals in a silicate melt. Solid particles influenced the final particle size distribution because they tended to remain intact during the experimental fragmentation process. An increasing proportion of alumina particles acted to increase the fragment size for a given decompression, suggesting that energy release was reducing. This would be consistent with the reduced proportion of gas phase that results from an increasing proportion of solids. This effect may also account for the observed increase in the decompression needed to fragment samples as the proportion of crystals increased, with the limit that a zero-vesicularity sample (100% crystals) would not fragment under decompression. Interestingly, tube bubbles were not formed in these experiments due to enhanced

heterogeneous bubble nucleation acting to reduce bubble diameter.

Spieler et al. (2004) carried out a systematic investigation of the rapid pressure drop required to cause the brittle spalling fragmentation of natural samples with a wide range of chemistry, porosity, permeability and crystallinity. The major control on fragmentation threshold pressure (ΔP_{frag}) was found to be the sample porosity (ϕ). Figure 10 shows the relationship to be approximated by $\Delta P_{\text{frag}} = 1/\phi$, with pressure given in MPa. Such a relationship makes qualitative sense because, as vesicularity tends to zero, ΔP_{frag} will tend to a large value because of little or no gas phase to drive the fragmentation. At large vesicularities, there will be little melt to resist the expansion of the volatile and fragmentation will take place with small decompressions. Figure 10 suggests that the rapid removal of order 100 m of overburden from a vesiculated magma with



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 10 Samples of natural volcanic material with a range of vesicularities were heated to 850 °C and decompressed in the ‘fragmentation bomb’ (Fig. 8) by Spieler et al. (2004). The pressure drop required to cause explosive brittle fragmentation was inversely related to the

sample vesicularity, with the constant of proportionality (here, about 1.07 MPa) being the tensile strength of the solid phase. Intriguingly, this suggests that heterogeneity in vesicularity will result in selective fragmentation of more vesicular regions of magma, particularly at vesicularities below 0.3. Data replotted from Spieler et al. (2004)

$\phi > 0.2$ will result in explosive fragmentation caused by pressure-retention in the pores. Such an event could happen by slope failure, for example, at Mt. St. Helens in 1980, or by a dome collapse. At lower vesicularities, the rapid removal of an overburden thickness of order 1000 m is required to initiate explosive fragmentation. However, although smaller overburden removals may not result in immediate spalling fragmentation, slower diffusively driven bubble growth will act to expand the magma and could thus trigger more sustained explosive activity.

Another mechanism of increasing pore pressure in magma is by crystallization of the silicate melt (e.g., Sparks 1997). This occurs because water is not incorporated into the structure of growing crystals and concentrates in the declining proportion of melt. Taddeucci et al. (2004) used the ‘fragmentation bomb’ to ascertain the fragmentation threshold of crystal-rich magma from the 2001 basalt erupted at Mt. Etna, as a function of porosity. This was combined with models of pressures generated by crystal growth, and calculations of conduit pressures required to eject magma blocks observed during the explosions. This approach supported an explosion model sourced in the heterogeneous fragmentation of crystallizing magma plugs. The particularly interesting aspect of this model is the introduction of porosity heterogeneity within the magma. Figure 10 demonstrates that as porosity increases, the pressure change required for fragmentation decreases. Therefore, the competence of a body of magma will depend on its weakest link, namely any regions of high porosity. Once these begin to fragment, then the less vesicular regions will be removed as large blocks.

Although volcanically accurate materials were used in fragmentation bomb experiments, the effects of reductions in length and timescale by three orders of magnitude remain unknown. Nevertheless, these important experiments give specific insight into key mechanisms that could operate during the rapid decompression (or internal pressurization) of a vesiculated body of magma, and general insight into the behavior of volcanic materials undergoing explosive eruption.

Explosion Products

Explosive expansion of magma, driven by the low bulk modulus of gas-phase volatile species, generates a range of consequent phenomena. The primary magma fragmentation (one mechanism is illustrated in Fig. 6, and another in Fig. 9) involves a large increase in surface area within the volcanic system. The mechanism of this increase is the breaking of molecular bonds.

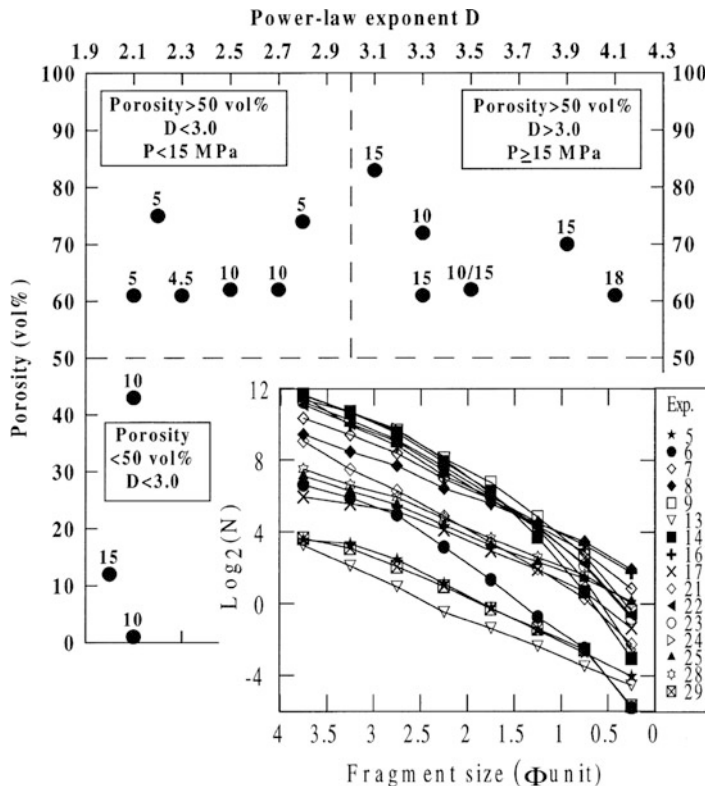
On exiting the volcanic vent, hot silicate fragments and volatile gas (mainly water vapor on Earth) interact with any atmosphere and cool. In the absence of a significant atmosphere, for example on Io, the volcanic ejecta expands, then the silicate fragments follow ballistic trajectories once they have decoupled from any volcanic gases and cool by radiation. Interaction between silicate fragments is now unlikely and their trajectories end when they impact Io’s surface; some may escape Io altogether. On Earth, the presence of an appreciable atmosphere results in different behavior. Explosion products emerge from the vent and interact with the atmosphere. If the silicate fragments are large and the explosive power low then fragment trajectories tend towards ballistic control. Conversely, if eruption power is high and fragments small, then atmospheric interaction may dominate. The mixing of air with hot eruption material and its consequent expansion may result in a thermally buoyant plume that ascends many kilometers into the atmosphere and may circumnavigate the globe by long-range atmospheric transport. If insufficient atmospheric air is entrained in order to develop and maintain buoyancy then some ejecta may fall back to ground to form a pyroclastic flow. On a smaller scale, the presence of electrostatic fields and liquid phases encourages sub-millimeter silicate fragments to aggregate. Many experimental studies have been carried out on the processes that follow explosive volcanic events and we review a selection here.

Fragment Size Distributions The fragment size distribution resulting from explosive volcanic eruption has been found to follow a power law distribution, $N = \lambda r^{-D}$, where N is the number of fragments greater or equal in size to r , and λ is a

constant of proportionality. For a single or primary fragmentation event the exponent (D) is predicted to be less than 3.0. Experiments carried out by Kaminski and Jaupart (1998) impacted a sample of Minoan pumice with a solid piston to give an exponent of 2.7. James et al. (2002) impacted two pumices together and found an exponent of 3.5. Pyroclastic flow or fall deposits from explosive volcanic eruptions exhibit an exponent range between 2.9 and 3.9.

Martel et al. (2000) rapidly decompressed hot vesiculated rhyolite in a 'fragmentation bomb'. Decompressions of less than 15 MPa produced exponents of less than 3.0 for the resulting

fragment size distribution, regardless of initial sample vesicularity. However, decompressions of >15 MPa, for samples of vesicularity >0.5 , resulted in exponents >3.0 and ranging up to 4.1 (Fig. 11). These results suggest that secondary fragmentation mechanisms, such as those induced by fragment-wall and fragment-fragment collision, could be occurring within the apparatus after the primary spalling-fragmentation event (Fig. 9). Alternatively, as high exponents were observed at high vesicularities, the highly heterogeneous nature of a foamed silicate could invalidate the assumptions made in suggesting 3.0 as the exponent boundary for primary fragmentation, at least for fragment size ranges smaller than the order of



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 11 Hot samples of bubbly rhyolite glass with a range of vesicularities were decompressed through various pressures in the 'fragmentation bomb' (Fig. 8), and the fragment size distribution (FSD) measured (Martel et al. 2000). The FSD was compared with a power-law distribution given by $N = \lambda r^{-D}$, where N is the number of fragments with radii larger than r . The power-law exponent, D , was found to increase with the magnitude of

decompression (numbers next to data points in MPa), as well as with higher vesicularity. Compare this to an exponent of 3.5 for colliding pumices in Fig. 12. Theoretically, $D > 3$ suggests more than one fragmentation process, but these are complex, heterogeneous, multiphase materials and established fragmentation theory based on dense homogeneous materials may not give the full picture. (Reprinted with kind permission from Elsevier)

bubble size. Interpretation of fragment size distributions between experimental and natural systems in terms of fragmentation processes is further complicated by difficulties in obtaining complete natural samples from explosive events. For example, winnowing in the atmosphere separates fragments as a function of fall velocity, which is a function of size and shape; therefore, the products of any individual fragmentation process are subsequently separated.

Fall Velocity of Silicate Fragments Understanding the behavior of silicate fragments in the atmosphere requires knowledge of the fall velocities of real volcanic pyroclasts, which rarely conform to an idealized spherical shape. Walker et al. (1971) dropped samples of pumice from heights up to 30 m and found that terminal fall velocities were best approximated by the fall equation for cylindrical bodies. Wilson and Huang (1979) carried out extensive experiments on silicate crystals and glass fragments in the size range 30–500 μm . A fragment shape factor ($F = (b + c)/2a$) was defined in terms of the longest (a), intermediate (b) and shortest (c) principal axes of the fragment, allowing definition of the empirical formula $Cd = (24/\text{Re})F^{-0.828} + 2(1.07 - F)^{0.5}$, where the fragment drag coefficient is given by Cd , as a function of Reynolds number, Re , and the shape factor. Suzuki (1983) revisited the experimental data and proposed a modified equation, $Cd = (24/\text{Re})F^{-0.32} + 2(1.07 - F)^{0.5}$. This empirical, experimentally informed approach allowed calculation of the fall velocity of a fragment in the atmosphere and provided input into numerical models of the dispersal of products from explosive volcanism. Further information can be found in Riley et al. (2003); Sparks et al. (1997); Textor et al. (2005).

Fragment Electrification James et al. (2000) investigated possible mechanisms by which silicate fragments could become electrically charged during explosive activity. Two natural pumices were collided together resulting in the generation of particles generally smaller than 70 μm in diameter as the silicate foam fragmented in brittle fashion. Fracto-emission generated ions and charged

silicate fragments. Ions were found to have one net charge and silicate fragments the opposite net charge. Experiments carried out at atmospheric pressure produced charge densities similar to those found on ash particles resulting from explosive terrestrial volcanism, whilst those carried out at 0.1 Pa sustained at least an order of magnitude more charge. The absolute amount of charge was dependent on the impact energy, (a result comparable to (Büttner et al. 1997) findings of increasing electric field with greater fragmentation energy), with the net charge representing a slight imbalance between positively and negatively charged fragments. The mechanism of producing charged silicate fragments from pumice-pumice collision adequately accounts for the fragment charges and electric fields measured in proximity to explosive volcanic eruptions. This suggests that pumice-pumice collision, i.e., secondary fragmentation, could be a major source mechanism for sub-100 μm silicate fragments.

Experimental hydrovolcanic explosions were found to produce an electrically charged fragment and ion cloud (Büttner et al. 2000). Interestingly, fragmentation using high-pressure Argon gas produced electrical effects of smaller magnitude than magma-water interaction, even though expansion rates of the ejecta from the experimental crucible were similar. This suggests that the cooling effect of water does produce much more brittle failure resulting in enhanced fracto-emission (Dickinson et al. 1988; Donaldson et al. 1988), consistent with the production of smaller blocky fragments from MFCI experiments (Zimanowski et al. 1997). It is also possible that polar water molecules dissociated into reactive ions during the MFCI process, a mechanism not so accessible to Argon atoms. Büttner et al. (2000) found striking similarities between experimental electric fields and those generated during explosive events at Stromboli volcano.

Silicate Fragments and Their Aggregation

Accretionary lapilli are widespread in the deposits from explosive volcanic eruptions, and anomalous variation in the deposit thickness and grain size distribution as a function of distance from the

explosive vent is commonly observed. Both of these processes are explainable if silicate fragments smaller than about 100 μm clump together, or aggregate, to form a larger entity which has different aerodynamic properties than its constituent particles. Sparks et al. (see Chap. 16 in Sparks et al. 1997) provides a review of particle aggregation in volcanic plumes prior to 1997.

Accretionary lapilli are obvious in the geologic record as millimeter to centimeter sized spheroidal aggregates (density range 1200–1600 kg m^{-3}) of silicate fragments and secondary minerals such as gypsum, and they formed the early focus of study into aggregation in volcanic plumes. Two complementary experimental studies provided insight into the formation mechanisms of accretionary lapilli. Gilbert and Lane (1994) focused on formation in atmospheric plumes. A wind tunnel was used to demonstrate experimentally that particles of volcanic ash collide and adhere to the surfaces of objects falling through the atmosphere that are covered with a thin layer of liquid. The liquid layer was found to act as a control on the sizes of silicate fragments that adhere, with high sticking efficiencies at 10 μm diameter, reducing by two orders of magnitude for fragments 100 μm in diameter. A key experimental finding was the importance of hygroscopic chemical species in maintaining thin liquid films. Experimentally this was achieved using NaCl; volcanically, sulphuric acid is the most likely hygroscopic compound that could maintain liquid films at low relative humidity and down to temperatures of -70°C . These experiments strongly suggest that accretionary lapilli are diagnostic of a three-phase environment within the volcanic plume in which they formed, supporting earlier hypotheses (e.g., Moore and Peck 1962).

Schumacher and Schmincke (1995) focused on accretionary lapilli formation in pyroclastic flow processes. Experiments were carried out providing insight into interpreting the structure of accretionary lapilli in terms of their formation mechanisms. Accretionary lapilli were synthesized by a method similar to the industrial production of fertilizer pellets, with mixtures of volcanic ash and variable proportions of water being

rotated in a steel pan. The resulting spheroidal agglomerates varied with the proportion of water present, formation being optimal at between 15 and 25% w/w water, and grain size sorting was found to occur at low water mass fractions. The mechanism of formation, namely capillary binding, is identical to that of Gilbert and Lane (1994), but under conditions of much higher fragment number-concentration.

The occurrence of secondary thickness maxima and bimodality of grain size distribution, exemplified by Carey and Sigurdsson (1982), implies the mechanism of silicate fragments being transported within aggregates. These aggregates could be accretionary lapilli, but these are often absent from the deposits suggesting that the aggregates are fragile and lose their identities once incorporated into a volcanic deposit. Fragile aggregates have been observed falling from volcanic plumes and one explanation for their formation is the presence of electrostatic charge. During fragmentation, silicate particles become electrostatically charged (see section “Fragment Electrification”), and whilst in the atmosphere these charges have no path to escape to earth unless electric fields exceed the level required for electrostatic discharge in air. This leads to long-lived attractive electrostatic forces between silicate fragments (Marshall et al. 2005), and consequently the formation of aggregates. Once on the ground an electrically conductive path to earth can be established, especially in the presence of moisture, and the aggregates disintegrate.

Schumacher and Schmincke (1995) carried out experiments using natural volcanic ash and a stream of ions from an electrostatic paint gun. Silicate fragments smaller than about 180 μm diameter were found to rapidly aggregate into loose fragile clusters demonstrating the viability of the electrostatic aggregation process for fine volcanic particles.

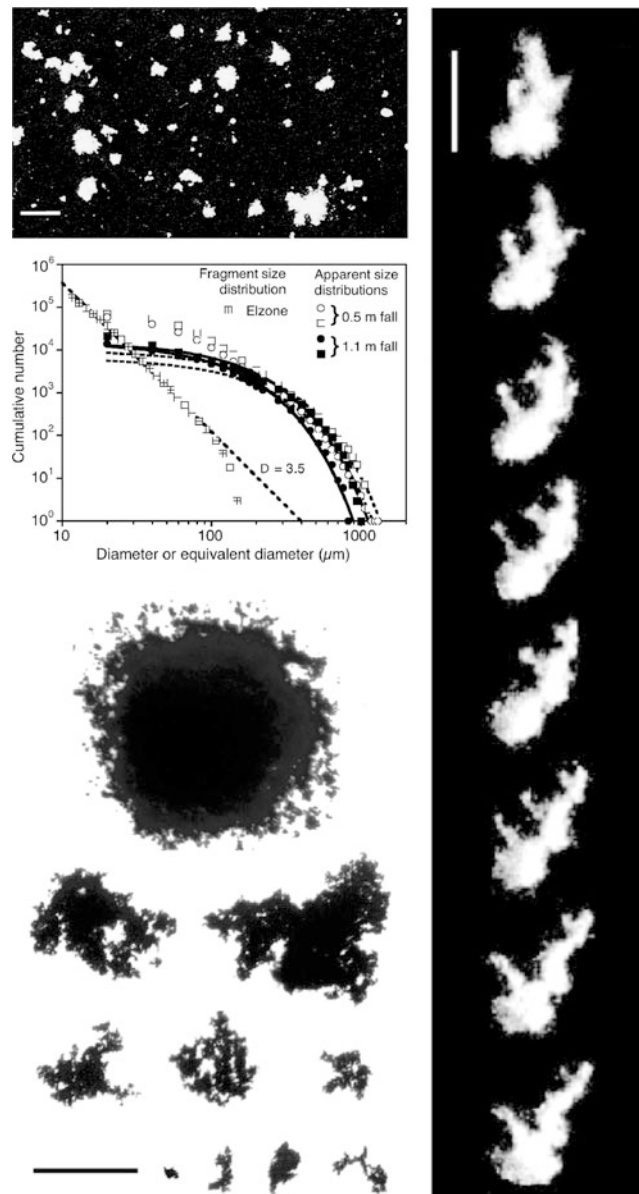
James et al. (2002) studied the generation of aggregates resulting from the interaction of ions and charged silicate fragments using the fragment generation and charging mechanism of James et al. (2000) with pumice from the May 1980, Mount St. Helens fall deposit. Silicate fragments were allowed to fall and interact over a distance of

0.5–1.1 m onto a variety of analysis platforms. Fallout was collected and analyzed for fragment size (Fig. 12) using an agitated dispersion of the particles in a conducting electrolyte to maintain disaggregation. The particle size distribution showed power law behavior with exponent 3.5 (Fig. 12), somewhat above the theoretical maximum of 3.0 for a primary fragmentation event. These experiments indicated that the aggregation process happens shortly after the fragmentation

event has generated charged silicate fragments and ions (Fig. 12). The aggregation process significantly changes the size distribution relevant to aerodynamic behavior (Fig. 12) from the single-fragment form $N = \lambda r^{-3.5}$ to aggregate form $N = 15,089 \exp[-0.011d']$, where d' is the aggregate diameter. Experimental measurement of aggregate fall velocities enabled estimation of aggregate densities between 80 and 200 kg m⁻³, at least an order of magnitude less than both the

Volcanic Eruptions, Explosive: Experimental Insights,

Fig. 12 Fragmentation during explosive eruptions produces electrically charged fragments and ions by fracto-emission processes (James et al. 2002, 2003). Electrostatic forces are significant in comparison to gravity for fragments less than about 100 μm in size, resulting in fragment aggregation. Aggregates have different aerodynamic properties than their constituent fragments, thus altering the fallout behavior of sub-100 μm volcanic ash and consequently its atmospheric transport. The top image shows experimental aggregates collected on a plate (scale 1 mm), with the graph showing the size distribution of the aggregates (curved lines), as well as that of the constituent fragments (straight line). The bottom panel (scale 0.5 mm) shows that experimental aggregates of different sizes have different morphologies, and the right-hand panel (scale 0.5 mm) shows strobe images of a falling and rotating experimental aggregate. (Reprinted with kind permission of AGU)



fragments that comprise aggregates and accretionary lapilli. The main consequence of electrostatic aggregation for sub-100 μm silicate fragments is to reduce atmospheric residence time because aggregation acts to increase fall velocity above the single fragment value. The size distribution of individual fragments within an aggregate was experimentally examined by James et al. (2003), who found empirically that for aggregate diameters less than 140 μm ,

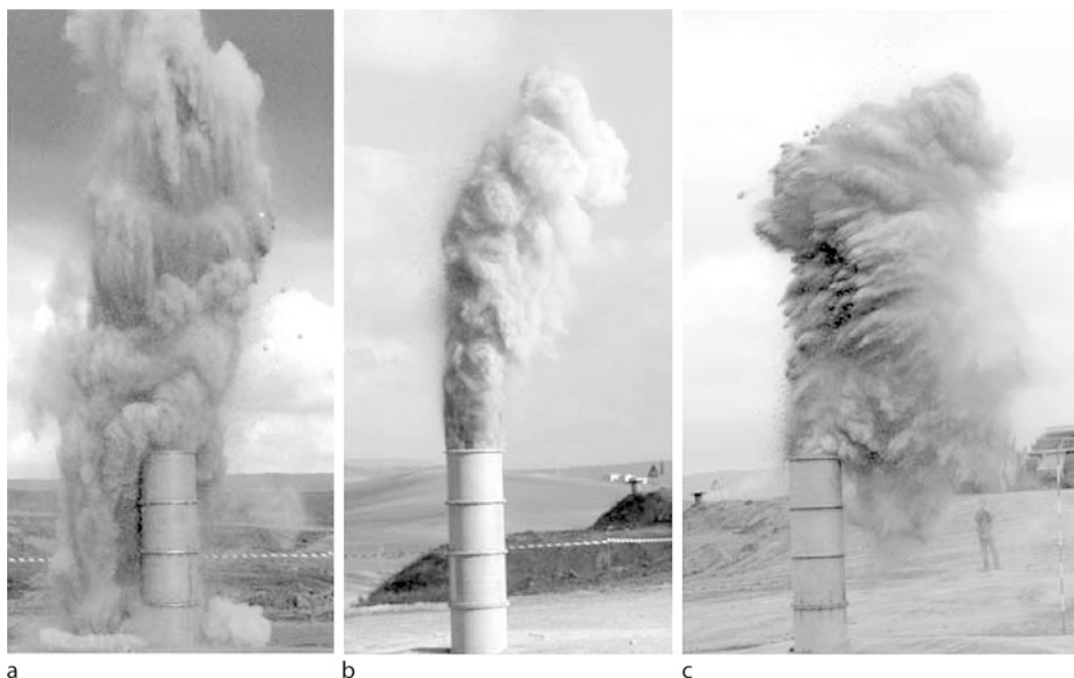
$$n = 0.000272d_a^3 \left[\exp(-0.22d_p) + 0.008 \exp y(-0.083d_p) \right],$$

where n is the number of fragments larger than diameter d_p within an aggregate of diameter d_a . As electrostatic aggregates exceed about 140 μm in diameter the size distribution of their constituent fragments was found to change. This was attributed to the aggregate growth mechanism becoming dominated by aggregate-aggregate interaction rather than aggregate-fragment interaction (Fig. 12), and stabilization of the size distribution of material being incorporated into the aggregate. Electrostatic aggregates can be represented as spheres of density about 200 kg m^{-3} and an experimentally determined empirical relationship between drag coefficient (C_d) and Reynolds number (Re) for aggregate diameter range 50–500 μm is $C_{d,\text{agg}} = 23\text{Re}^{-0.637}$ (James et al. 2003). Such experimental data can then be used within numerical models of the transport of sub-100 μm volcanic ejecta in the atmosphere, although at the time of writing this has yet to be undertaken.

Pyroclastic Flows Pyroclastic flows comprise a hot mixture of silicate particles and gas formed by the collapse of negatively buoyant regions of a volcanic plume or avalanching of material from a growing lava dome. By nature, pyroclastic flows are destructive to infrastructure and almost invariably fatal to plants and animals. Insight into the behavior of pyroclastic flows may be gained by studying their deposits. However, it is difficult to definitively identify fluid dynamic processes operating during a pyroclastic flow from the complex structures within deposits that are revealed once

motion has ceased. Here, experiments can help in isolating flow processes and their subsequent effect on deposits. However, scaling between experiment and natural process requires care to ensure behavioral similarity as particles of a range of sizes interact with the gas phase to different degrees.

Dellino et al. (2007) designed experiments to investigate the mechanics of pyroclastic flows on a scale comparable to the natural process. In order to approach this, the physical scale of the experiments was such that they required an outdoor laboratory. The experimental material was sourced from a natural pyroclastic flow deposit in order to mimic the interaction between fragments and gas as closely as possible. Varying masses of this material were packed into an experimental tube 60 cm in diameter and 2.2 m high. 14 liters of gas at pressures over 10^7 Pa was rapidly injected into the base of the tube to eject the experimental material, with experiments triggered and logged under computer control. The experiments were carried out at ambient temperature and were, therefore, not scaled for any thermal effects; this has the benefit of reducing both equipment and process complexity. Ejected material could be sampled for analysis and compared to natural pyroclastic deposits. Preliminary experiments showed two main types of behavior as the specific mechanical energy (SME, given by *gas pressure $\times$ gas volume $\div$ mass of pyroclastic material*) of the system was changed (Fig. 13). At high SME values a dilute particle plume develops with the particles settling out individually. At low SME values the ejected material forms a collapsing column that generates a particle flow similar to a pyroclastic flow. Intermediate SME values give transitional behavior with aspects of both end-members. These experimental observations have considerable phenomenological similarity to natural explosive eruptions where particle-dilute regions ascend in the atmosphere and more particle-concentrated regions fall under gravity to give pyroclastic flows. Such similarity gives added confidence that mechanisms operating in the natural process also operate in the simulation.



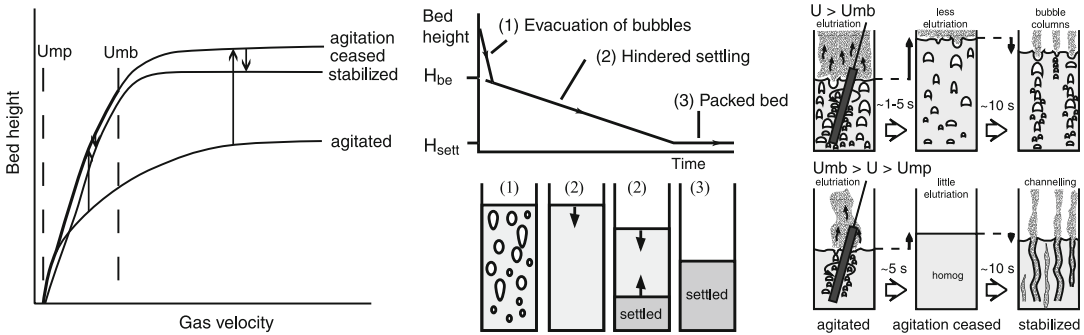
Volcanic Eruptions, Explosive: Experimental Insights, Fig. 13 Ejection of natural pyroclastic material by a burst of gas can be used to study both volcanic plumes and pyroclastic flows (Dellino et al. 2007), this also looks enormous fun! a shows an experiment with low specific mechanical energy (SME) generating a collapsing column

and (pyro)clastic flow. c shows a high SME experiment generates a plume that, whilst not buoyant, disperses on the prevailing wind, and b shows the transitional case. Such large-scale investigation helps to overcome some of the scaling difficulties imposed by smaller-scale laboratory experiments. (Reprinted with kind permission of AGU)

Experiments producing a collapsing-column deposited material at the point of impact of the column with the ground. The resulting deposit was found to be structureless, just as many natural proximal pyroclastic deposits, because grain-sorting processes did not operate in this concentrated particle flow. Other similarities included the formation of a concentrated undercurrent, bed load at the base of the turbulent flow, and continuous atmospheric suspension of small silicate particles; this is very interesting in the absence of thermal processes. These preliminary experiments demonstrate great potential for linking processes in scaled experimental flows with the resulting experimental deposits and, therefore, providing a powerful interpretation tool for natural pyroclastic flows.

Experiments carried out at smaller scale, but which included heating of the volcanic ash (Druitt et al. 2007), have also investigated gas retention in

pyroclastic flows. A bed of pyroclastic flow material with a wide fragment size range was fluidized by the drag force of rising hot gas. Both the bed expansion and collapse (degassing) after the gas supply was cut were studied. During expansion, smaller fragment size, lower density and higher temperature all promoted more uniform and easier expansion of the fluidizing bed. Figure 14 illustrates the nature of flow patterns that develop. Collapse experiments were imaged using X-rays and carried out between 20 °C and 550 °C giving temperatures comparable to those in natural flows. The generic collapse process is illustrated in Fig. 14. Experiments and numerical models identified that an aerated bed loses its gas by permeable flow on a diffusive timescale. A fluidized and uniformly expanded bed was controlled by time-scales of hindered settling and diffusive degassing of the resulting sediment layer. The relative magnitudes of these timescales depended on the



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 14 The inflation and deflation of a bed of natural pyroclastic material using hot gas shows a range of flow structures (Druitt et al. 2007). During inflation, the bed can be homogeneous or develop gas bubbles or channels, and fine particles can be removed by elutriation. When fluidizing gas velocity exceeds U_{mp} (left panel), the bed inflates and particles start to segregate on the basis of size and density. Bubbling occurs when gas velocity exceeds U_{mb} (upper right panel), and the bed expands only weakly with increasing gas velocity (left panel). Agitation of the flow acts to reduce the gas volume fraction.

Once the gas supply is closed off then bubble collapse results in rapid deflation ((1) in middle panel). An extended period of hindered settling then generates a layer of settled material at the base of the bed ((2–3) in middle panel). These experiments were used to investigate the gas retention properties of pyroclastic materials as a function of fragment size and temperature. Gas retention was favored by finer fragment sizes, with high temperature having a secondary retention effect. Hot, fine-grained pyroclastic flows are, therefore, predicted to flow further. (Reproduced with kind permission from Springer Science and Business Media)

thickness of the degassing bed, with settling dominating in thin beds and diffusion in thick ones. These experiments indicate that hot pyroclastic flows with small fragment sizes retain gas longer than cool, large fragment flows, with temperature having a secondary effect. The gas retention translates into the distance a flow can travel and these experimental results are consistent with observations of natural pyroclastic flows.

Analogue Approach

Experiments with volcanic materials can suffer constraints placed by the high temperatures and pressures at which the processes of interest occur. If similar processes can be studied under more amenable environments, then the practical limitations are relaxed and, generally, fluid volumes can be increased and a greater range of transducers are available for measurement. This can be achieved by substituting natural fluids with analogues that can be studied at near room temperatures and pressures. However, one difficulty with this approach is in ensuring the relevance of analogue results to volcanic systems and, perversely,

natural volcanic materials are now often characterized more fully than many of the analogue fluids used.

Explosive Processes

Direct observation of the motion of magma within volcanic conduits would reveal the how and why of volcanic explosive activity. However, direct observation is considered near-impossible and indirect observation open to interpretation. Experiments with natural materials provide constraints on physico-chemical processes within conduits, but extending the experimental parameter space by using analogue materials provides greater insight and enables testing of numerical models.

High VEI Events Volcanic eruptions of high VEI involve the transition of a low-velocity, high-density, high-viscosity bubbly magma, for which the liquid phase is considered continuous, to a high-velocity, low-density, low-viscosity ‘dusty gas’ where the gas phase is considered continuous. This fragmentation process represents the point at which the liquid and gas phases start to separate. Studies designed to investigate this process in general, but with volcanic

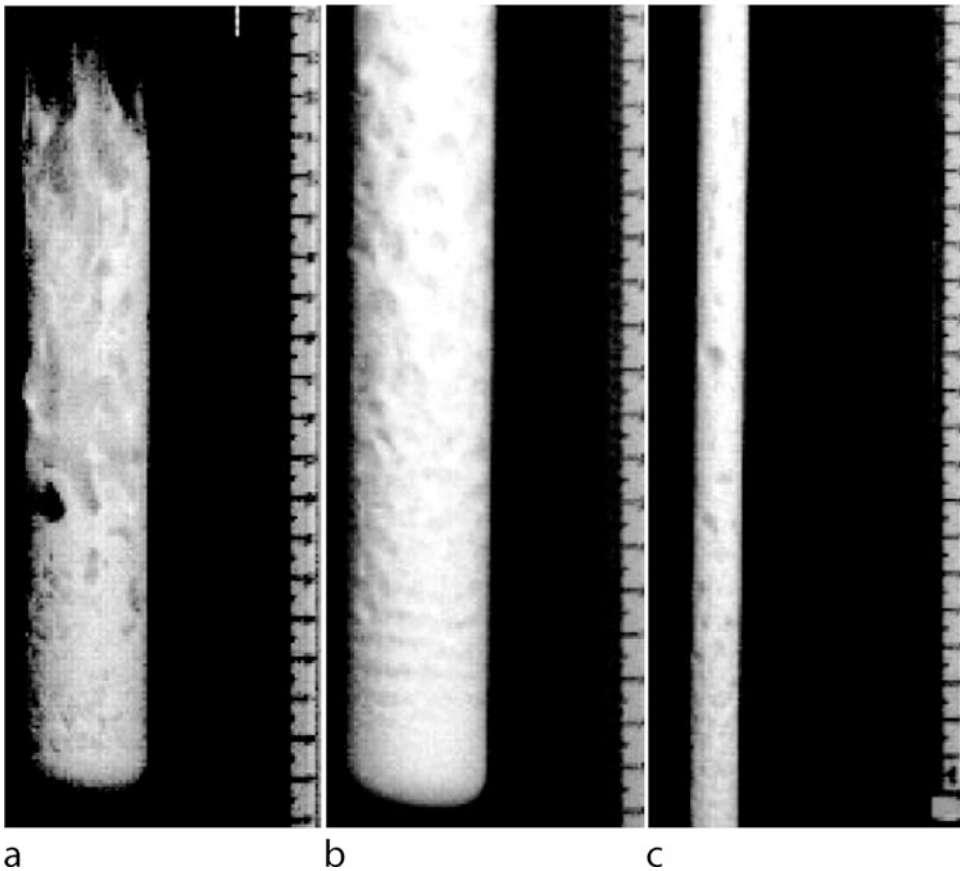
application in mind, form a major part of experimental research using the analogue approach.

As pressure declines, the exsolution and expansion of volatile (specifically water) from magma provides the driving mechanism for most explosive events. In order to study similar behavior in analogue systems the shock tube has been almost universally adopted, as it has for the investigation of fragmentation in natural materials (e.g., Alidibirov and Dingwell 1996a, b). Such experiments became of great importance around WW2 when understanding of shock waves and flame fronts from explosions, and supersonic flight, became of interest to the fluid dynamic community; volcanological application of shock tubes emerged from this background. Mader (1998) provides a review of early shock tube experimentation relating to flow within volcanic conduits, we briefly summarize these here, but focus on more recent work.

Post-Fragmentation Flows The decompression of beds of small, incompressible particles with compressible, gas-filled pore space can give insight into the development of two-phase flows within and above the conduit subsequent to primary fragmentation (Anilkumar et al. 1993). The resulting rapidly expanding, but transient flows show development of significant heterogeneity, which declines as expansion reduces particle concentration. The development of heterogeneity within expanding particle-laden gas suggests mechanisms for the formation of pyroclastic flows from high-density sections of eruption columns, whilst lower-density sections become buoyant and form an eruption plume. Using smaller particles, Cagnoli et al. (2002) carried out similar experiments to explore the dynamics of short-lived Vulcanian explosions. In similarity (we think, but see Fig. 9) with post-fragmentation volcanic flows, the experimental tube was much larger than the particle size. As in Anilkumar et al. (1993), these transient flows demonstrated heterogeneous distribution of particles as the mixture expanded and the gas separated from the particles (Fig. 15). These features include sub-horizontal gas-rich regions that invite comparison with those found by Spieler et al. (2003) in decompressed

pumice samples; the implication being that the primary fragmentation process may be incidental to the nature of the evolving flow. Heterogeneities also comprise gas-rich bubbles, sometimes approaching tube diameter in size, and more elongated towards the flow front. Streamers of particles also emerge from the flow front indicating inhomogeneity in the distribution of gas escaping the particle flow (Fig. 15). As the flow ages and velocities increase, turbulence acts to smear out heterogeneities. The source of the heterogeneities in such transient flows is unclear. They may represent the expansion of gas rich regions in the original packed bed, suggesting that the transient flow reflects heterogeneity in the source material. Variability in permeability, and the dynamics of initial particle acceleration may also act to amplify heterogeneity as the flow expands. These experiments demonstrate the complex nature of even ‘simple’ two-phase flows that are undergoing rapid changes in variables such as pressure and velocity. The volcanic implication is that flow within a volcanic conduit can be highly heterogeneous and that modeling such flows as having smoothly changing density, as a function of time or space, will be far from representing the real richness of the flow physics. Such modeling may give accurate prediction of time-averaged parameters, but will not reflect variability. However, it could be the variability that ultimately results in hazards like pyroclastic flows.

Chojnicki et al. (2006) demonstrated that models of explosive volcanic flows based on pseudogas approximations underestimated the initial shock wave strength and velocity. Pseudogas models, and others based on steady-state experiments, then overestimated the subsequent particle bed expansion rate when applied to laboratory scale experiments. Such discrepancies suggest that one or both of the approaches, numerical and experimental, requires modification to reconcile the differences in order to understand the natural phenomena. In this case, Chojnicki et al. (2006) identify processes not accounted for in the numerical approach, but which became apparent from experimental observation. Existing theory was then empirically modified to fit the unsteady



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 15 Rapid 20–90 kPa decompression of beds of sub-100 μm glass beads generates a particle flow as the gas phase expands in response to the pressure drop (Cagnoli et al. 2002). a shows the decompression of 100 g of beads (with average diameter of 38 μm through 51 kPa) after 28 ms. Note the wispy *flow front*, indicating escape of gas, and the development of large gas bubbles within the

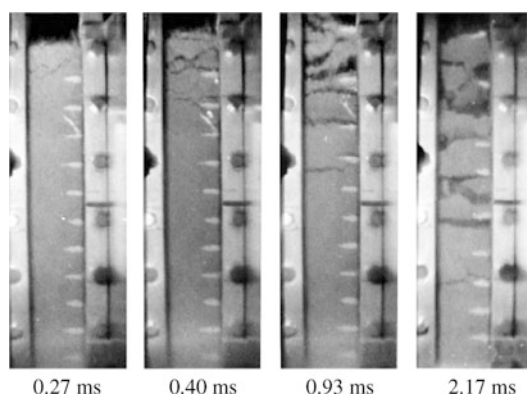
flow (scale shows 1 cm and 0.5 cm gradations). b (150 g of 95 μm beads, 69 kPa decompression, after 23 ms) shows the development of sub-horizontal particle-poor regions, giving some phenomenological similarity to spalling fragmentation (Fig. 9). c shows flow in a narrower tube (100 g of 38 μm beads, 68 kPa decompression, after 20 ms) where large particle-poor regions develop at the flow margin. (Reprinted with kind permission from Elsevier)

experimental flow. Volcanically, these experiments suggest that the pressures present in conduits prior to Vulcanian explosive events were underestimated by a factor of approximately five when calculated using established pseudogas and inviscid shock theories. The pressures calculated from equations based on volcanically relevant experiments are consistent with the overpressures associated with Vulcanian explosions, as well as the rupture strength of volcanic rock. This demonstrates the importance of experimental design to progressing understanding of complex volcanic phenomena, and using the results of these

experiments to test, modify and develop the numerical approach.

Decompression-Driven Flows The rapid decompression of natural vesicular solids causes fragmentation if the pressure drop is large enough for pressure gradients within the sample to exceed the tensile strength (Fig. 9). Such transient experiments have also been carried out on analogue solids in order to further characterize the behavior of this system of flows. Decompression of synthetic organic resin, with about 90% porosity and significant permeability (Alidibirov and Panov

1998), resulted in fragmentation phenomenologically similar to that of natural samples (Figs. 9 and 16) undergoing brittle failure. A fragmentation front or wave was observed to propagate through the sample creating spalling-type fragmentation. Fragment size was observed to decrease, and exit velocity increase as the magnitude of decompression increases from 0.1 to 0.8 MPa. The propagation velocity of the observed fragmentation wave was less than sound speed in either the solid resin or the gas in the pores. Combining this observation with the dependence of fragment size on pressure drop suggests that fragmentation occurs as gas escapes from the decompressed foam surface, creating a pressure gradient within the foam. When this pressure gradient is sufficiently steep, the foam fails revealing the next gas-escape surface. The expanding gas then accelerates the foam fragments. This implies that fragment size will also depend on sample permeability; an impermeable sample fragmenting on a scale comparable with bubble size and at small decompressions, whilst a permeable sample requires large decompressions and fragments on a scale much larger than bubble size. Very permeable materials will not fragment because the gas can escape without applying sufficient tensile force to the foam.

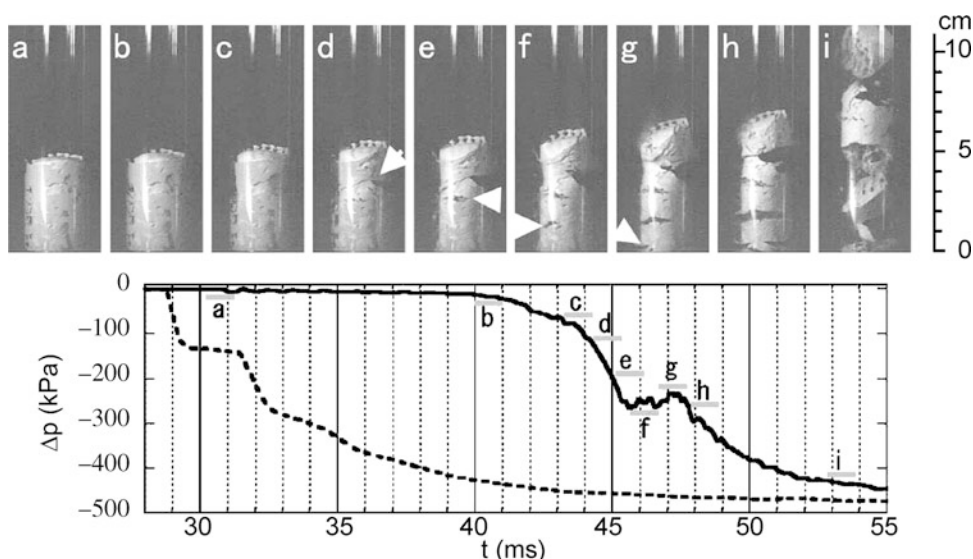


Volcanic Eruptions, Explosive: Experimental Insights, Fig. 16 An 0.8-MPa rapid decompression of vesiculated solid foam (Plastiprin) shows very similar spalling fragmentation behavior to natural vesiculated magma (compare with Fig. 9). This demonstrates the usefulness of analogue experimentation in identifying the universality (or not) of flow behaviors (Alidibirov and Panov 1998). (Reproduced with kind permission from Springer Science and Business Media)

Variability in permeability and porosity may well be reflected in fragment size for any decompression event.

Ichihara et al. (2002) depressurized a vesiculated silicone compound and studied the expansion of vesicular materials across the viscoelastic transition by varying the decompression rate (Fig. 17). Fragmentation occurred at decompression rates $>2.7 \text{ MPa s}^{-1}$, with only sample expansion occurring at $<2.7 \text{ MPa s}^{-1}$. At high decompression rates, fragmentation occurred before expansion of the fragments. These experiments showed that both the fragmentation threshold and the nature of the fragmentation process depend not on the magnitude of decompression (above a minimum), but on the rate of pressure change. Furthermore, the rate of decompression needed to exceed the critical value for more than about 0.1 s (Ichihara et al. 2002), indicating that a pressure change approaching 300 kPa represents the minimum pressure drop for fragmentation and also relates to the tensile strength of the experimental material. This fragmentation timescale was similar to the structural relaxation time for the glass transition (about 0.3 s), suggesting that crossing the relaxation timescale results in brittle failure of the sample. Coincidentally, the timescale for viscous control of bubble growth in the experimental material is also in the region of 0.1 s, suggesting that bubble expansion may determine the fragmentation timescale. This coincidence occurs because the rigidity of the silicone fluid is within an order of magnitude of the experimental pressure; in magmas the difference is in the region of three orders of magnitude. The nature of the fracture surfaces (Fig. 17, compare with Figs. 9 and 16), and the observation of minimal expansion before fragmentation suggests, however, that fragmentation occurred because instantaneous strain rate exceeded that required for brittle failure.

Using viscoelastic and shear-thinning solutions of 0.1%, 0.3% and 0.5% w/w polysaccharide (xanthan gum) in water, transient flow experiments exploring a wider parameter range, including variable porosity, were undertaken by Namiki and Manga (2005). Air bubbles were introduced



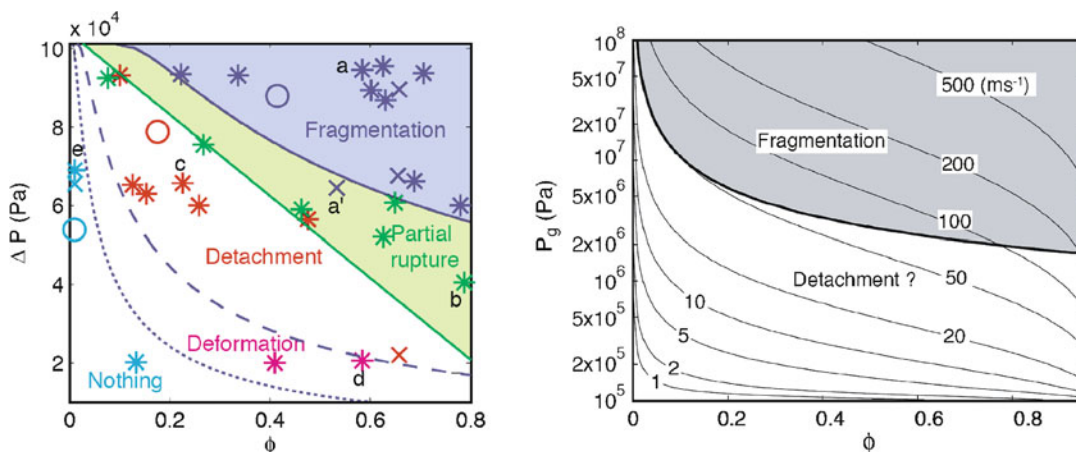
Volcanic Eruptions, Explosive: Experimental Insights, Fig. 17 Rapid (50 MPa s^{-1}) 0.5-MPa decompression of a vesiculated dilatant (*shear-thickening*) silicone compound that acts as a brittle solid at high strain rates ($>3 \text{ s}^{-1}$), but a viscous liquid at low strain rates. Pressure above the sample is given by the dashed line, whilst that below by the solid line. The surface of the sample is decompressing in (a), but pressure beneath the

sample remains steady until (b), decreasing more rapidly from (c). Brittle failure of the sample occurs over a $<10 \text{ ms}$ time span (d–g) in a spalling-fragmentation manner (Ichihara et al. 2002), about 40 ms after initial decompression. Fractures appear to initiate at the tube wall and then open to expand the flow (h, i). There is little or no expansion of the vesiculated sample or fragments. (Reprinted with kind permission of AGU)

using a hand-mixer, giving gas volume fractions up to 0.79. Samples were rapidly decompressed from atmospheric pressure to pressures ranging from 0.8 to 0.06 of an atmosphere; pressures much lower than this would encounter the saturated vapor pressure of water, causing the flows to be influenced by the production of water vapor. The shock tube was 0.05 m diameter by 0.25 m length, with fluid added to a depth of about 0.05 m. A range of flow patterns was observed (Fig. 18) as system parameters were varied. These include: fragmentation, partial rupture, detachment, deformation, and nothing; only ‘fragmentation’ results in ejection of fluid from the shock tube. Large decompression of highly vesicular samples leads to fragmenting flows (Fig. 18), which initially expand rapidly, then fragment layer by layer (Fig. 19) showing some similarity to spalling fragmentation (Figs. 9, 16 and 17). In ‘partial rupture’, large-scale net-like structures develop that allow gas to escape from the fluid, but the liquid phase does not fragment *en masse*. At lower flow energies, the liquid phase detaches

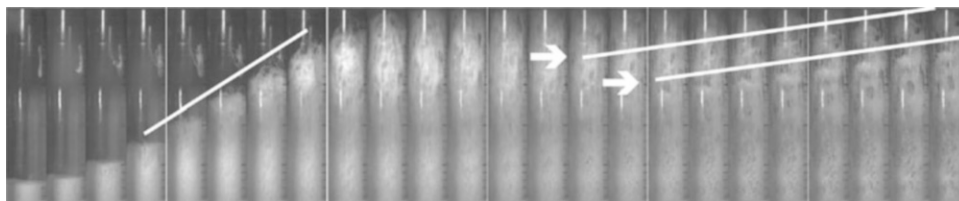
and contracts from the tube wall to yield an inverse annular, or ‘detached’ flow. Transient experiments allow the measurement of flow-front position, thus enabling estimation of strain rates within the flow. In these experiments the flow expansion was found to be independent of viscosity, which ranged from about 30 Pa s to 0.01 Pa s and was strain-rate dependent. For all experiments, it was found that shear rates were sufficient for elastic behavior, consistent with the independence from viscosity.

The changes in flow pattern are attributable to the behavior of the bubble walls. Detachment of the fluid from the tube walls can be explained by the Poisson’s ratio of the axially expanding foam, and represents the development of radial heterogeneity within an axial flow. However, if bubble walls rupture, elastic energy is released resulting in radial expansion and, thus, suppression of detachment. This suggests that fragmentation cannot take place from the detached flow pattern. Failure of bubble walls also increases permeability and allows gas to escape. Failure of the thin



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 18 Aqueous xanthan gum solutions are shear thinning and develop brittle solid behavior at strain rates above about $0.1\text{--}1\text{ s}^{-1}$, depending on concentration. The left panel shows that rapid decompressions ($<0.1\text{ MPa}$) of pre-vesiculated xanthan gum solution results in a range of flow patterns (Namiki and Manga 2005). Combining data established from the

decompression of natural samples with the analogue xanthan gum experiments allows the calculated fragmentation field for explosive volcanic eruptions (right panel) in terms of gas pressure and vesicularity. It was not possible to define the parameter space for other experimental flow patterns applied to magmatic systems. (Reprinted with kind permission from Elsevier)



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 19 Fragmentation of vesiculated xanthan gum solutions occurs if decompression and vesicularity are high enough (Fig. 18). The first frame (zero time) shows the sample of vesiculated solution. On decompression, bubbles expand to give an approximately constant-velocity flow front (Namiki and Manga 2005), shown by the line in

frames 4–8 (3 ms between frames). During expansion, bubbles coalesce through film breakage to increase significantly in size, especially in the upper parts of the flow. The foam then fragments layer by layer, as shown by the two lines following liquid-poor regions. Most of the xanthan gum is ejected from the shock tube during the experiment. (Reprinted with kind permission from Elsevier)

film, or plateau, between two bubbles, but not of the thicker plateau borders between three or more bubbles, is considered to result in the partial-rupture flow pattern. The plateau borders form the net-like structure. Wholesale failure of plateau and plateau borders results in fragmentation. Application of experimental processes to volcanic scales (Fig. 18) by utilizing results from studies of transient flows of natural materials, gives indication of the decompressions needed to fragment vesicular magma for a range of vesicularities,

but defining other flow patterns at volcanic scale requires further research.

Using a similar technique to Ichihara et al. (2002), Namiki and Manga (2006) investigated the role of transient decompression rate on aqueous xanthan gum solution. The variables here also included porosity ($0.19\text{--}0.82\text{ v/v}$) and absolute decompression ($0.41\text{--}0.94\text{ atmosphere}$). High decompression rates result in a variety of flow patterns, including fragmentation, depending on porosity and absolute decompression. At low

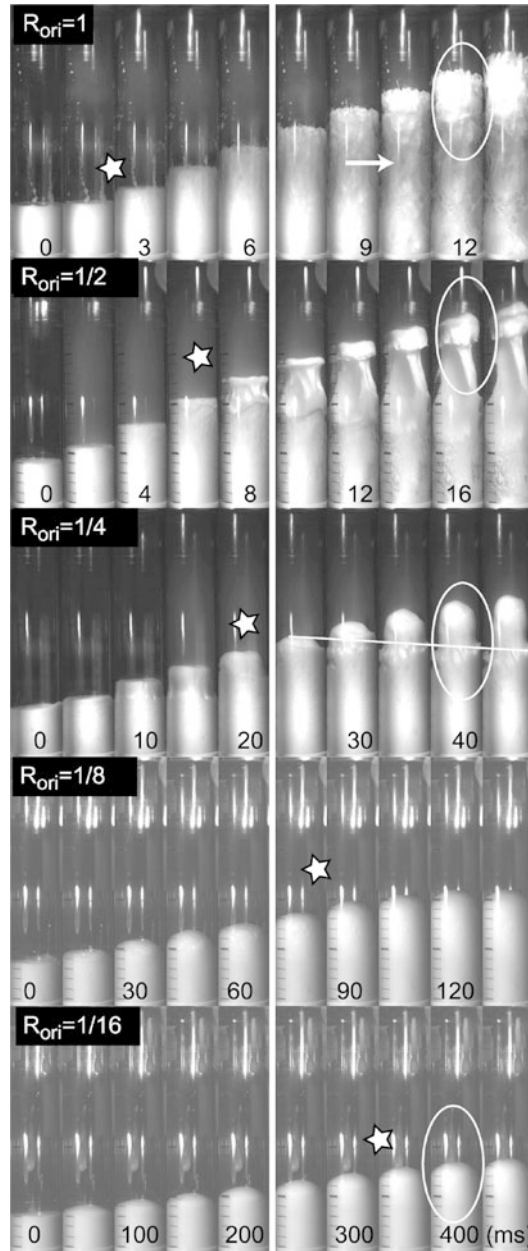
decompression rates only sample deformation or expansion takes place (Fig. 20). Rough upper surfaces develop on flows during high decompression rates, indicating that bubble walls are rupturing; conversely, lower rates of decompression leave flows with smooth upper surfaces. These experiments suggest that decompression rate is an important parameter in determining eruption behavior, in particular the transition between effusive and explosive eruption in low-viscosity vesiculated magmas subjected to decompression. Explosive response (fragmentation) to decompression results from disequilibrium expansion, where pressure in the bubbles is significantly higher than in the surrounding fluid and bubble expansion is limited by the enthalpy change resulting from the decompression. Bubbles cannot physically expand fast enough to maintain the same pressure as the surrounding fluid. Effusive response results from bubble pressure being close to fluid pressure.

Models of equilibrium and disequilibrium expansion were tested against experimental results. At high decompressions, equilibrium expansion would be faster than disequilibrium expansion; experiments follow the slower disequilibrium expansion model. At lower decompressions the disequilibrium model predicts more rapid expansion than the equilibrium model; experiments again follow the slower expansion, this time in equilibrium. Namiki and Manga (2006) gave the critical decompression rate for inviscid explosive behavior as $-P'_{Ot} > (2\gamma/\rho_L \phi_i P_{Gi}(1 - \phi_i)(\gamma - 1))^{0.5} (P_{Ot}^2/h_{Fi})$, where P_{Ot} is pressure outside bubble during decompression, γ is isentropic exponent, ρ_L is liquid density, ϕ_i is initial gas volume fraction, P_{Gi} is internal bubble pressure and h_{Fi} the initial height of the bubbly fluid column. This experimentally based threshold decompression rate is consistent with the estimated decompression rate for the explosive/effusive transition in natural basaltic eruptions (Fig. 21), with Hawaiian eruptions suggested as mainly driven by the sufficiently rapid decompression of bubbly magma.

Degassing-Driven Flows The decompression and expansion of materials with pre-existing

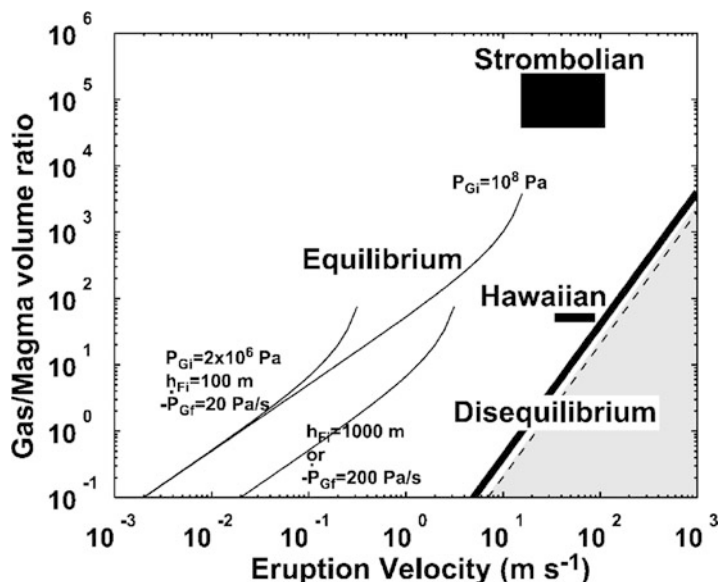
vesicularity provides valuable insight into volcanic behavior. However, magmas also degas on decompression, potentially adding to the explosivity of eruptions. In contrast to experiments with natural materials, where high temperatures and pressures make spatially and temporally extensive flows difficult to achieve, analogue experiments offer the possibility of studying the evolution of flow processes from single phase liquid to multi-phase fragmented flow using high-speed imaging and other data collection. Decompression of an unvesiculated hydrated magma results in, amongst other effects, the nucleation and growth of bubbles of gas phase or supercritical water, complex changes in the rheology of the liquid phase, increasing permeability and decreasing diffusivity. Incorporating all these interdependent variables into an experimental system adds complexity and makes understanding the first order effects more difficult, even if scaling between experiment and volcano is possible. Designing experiments to investigate the dominant physics using a canonical approach reduces complexity and eases scaling requirements. The first order physical process occurring during these flows is the phase change of one system component from liquid (or more strictly, dissolved) to gas, with the associated reduction in bulk modulus and response to pressure change. Therefore, the aim of analogue experiments here is primarily to study the effect of degassing on flow.

The decompression of a liquid below its saturated vapor pressure results in boiling of that liquid as it becomes superheated. Experiments carried out by Hill and Sturtevant (1990) used a shock tube to rapidly decompress a refrigerant. The refrigerant was Newtonian in rheology and of low viscosity, which, whilst dissimilar to the physical properties of magma, reduces the experiments to the first order process. Boiling produced a flow of vapor with entrained liquid drops. The superheating of the experimental liquid can be likened to the supersaturation of water in magma, and the liquid boiling to the exsolution of water from magma. The emerging gas phase expands, entrains liquid droplets and rapidly



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 20 Images of a 60-kPa decompression of xanthan gum solution of initial vesicularity about 0.6 (Namiki and Manga 2006). In order to study the effect of decompression rate, the top of the shock tube has been choked with an orifice whose radius is a fraction of the tubediameter (R_{ori}). Experiments with rapid decompressions have short durations (time in ms on individual frames), and slow decompressions significantly longer durations, with the star indicating the time decompression ended. For $R_{\text{ori}} = 1$, the flow develops a rough surface

suggesting rupturing of bubble walls (ellipse), and the arrow indicates the onset of fragmentation. When $R_{\text{ori}} = 1/2$, the flow top starts to separate (ellipse), but surfaces are smooth indicating that bubble walls are staying intact. Fragmentation does not occur when $R_{\text{ori}} = 1/4$, but the flow detaches from the tube wall as it expands. Smaller orifices result in flow expansion that retains wall contact. Different values of vesicularity and absolute decompression yield different patterns of behavior. (Reprinted with kind permission of AGU)



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 21 Application of experiments from Fig. 20 to the volcanic case give an estimate of the magma velocity required to cause sufficient decompression rate to initiate explosive eruption (gray region) as a function of magma vesicularity (Namiki and Manga 2006), where P_{Gi} is initial bubble gas pressure, h_{Fi} initial height of bubble column and $-P'_{Gf}$ the bubble decompression rate.

Eruptions that result from the rapid expansion of bubbly magma are likely to plot on the thick line separating equilibrium and disequilibrium fields. Hawaiian eruptions approach the transition and, on this basis, result from the rapid expansion of vesiculated basalt magma. Strombolian eruptions, however, are indicated to result from another mechanism. (Reprinted with kind permission of AGU)

accelerates the flow. These experiments showed the development of a sharply defined plane at which explosive boiling occurred.

The addition of inert particles to a volatile liquid (Sugioka and Bursik 1995) reduces the proportion of volatile phase and maintains particles in the post-fragmentation flow. Decompression, combined with the imposition of a steep pressure gradient within a centrifuge, results in a well-defined fragmentation front when decompressions are large. However, the mass of inert material acts to reduce post-fragmentation acceleration thus maintaining pressure and suppressing heterogeneous bubble growth below the fragmentation front. Small decompressions result in bubble nucleation on the inert particles throughout the mixture, illustrating that system behavior depends on the rates at which processes respond to imposed changes. Although these experiments probably represent end-member behavior, they provided a tantalizing glimpse of possible

explosive fragmentation mechanisms that may operate in volcanic conduits given suitable conditions.

Volcanically, only a few weight percent of the magma degasses to drive the flow, whilst boiling experiments represent an extreme expression of degassing flows where all the liquid can change phase to gas. Experiments that retain low liquid viscosity and Newtonian rheology, but degas only a fraction of their mass, extend our knowledge of degassing flows and are likely to behave in a different, possibly more volcanic fashion. Mader et al. (1994, 1996) rapidly generated bubbles of CO_2 in water, and throughout the fluid rather than just at the top interface. In volcanic systems, if bubbles nucleate and grow on timescales over which they can rise and escape then significant fluid expansion does not occur. However, if the CO_2 can be experimentally generated with sufficient supersaturation to drive bubble growth on timescales much shorter than those needed for

bubble rise, then the two-phase mixture rapidly expands and fragments in a ductile fashion, producing liquid and foam drops suspended in gas, as the gas volume fraction increases. Two methods were used to achieve CO_2 supersaturation up to 455 times ambient. Water was saturated in CO_2 and decompressed in a shock tube developing high supersaturations. An alternative approach used rapid chemical reaction between K_2CO_3 and HCl to generate a supersaturated CO_2 solution. An intriguing positive feedback mechanism was observed whereby agitation of the liquid by the growing gas bubbles enhanced the release of gas thus increasing the violence of the flow. It was found that considerable expansion and acceleration took place before the major change in flow pattern induced by fragmentation. This contrasts with brittle systems that tend to fragment then expand (Ichihara et al. 2002). Accelerations approached 180 g and were proportional to volatile supersaturation; velocities peaked at 15 m s^{-1} , and strain rates were in the region of 30 s^{-1} . These are considered to be in the same range to those experienced volcanically and indicate a degree of similarity between laboratory and natural systems, with strain rates suggesting that high-viscosity magmas will be fragmenting in brittle fashion.

Injection apparatus in the chemical mixing experiments introduces discontinuities in the tube geometry and this appears to play a role in the behavior of the flow. Mader et al. (1997) carried out relatively sustained explosive experiments, lasting about 1.5 s , where a considerable volume of carbonated water was decompressed. The vessel used was a round-bottomed flask. The discharge rate of water was found to fluctuate because of flow instabilities at the base of the flask neck, probably accompanied by pressure fluctuations. The flask neck was also found to pin the point of fragmentation as the flow evolved, with similar results being found by Zhang (1998). These experiments demonstrated the importance of sustained flows in allowing quasi-steady behavior to emerge, as well as demonstrating that conduit geometry influences flow patterns.

Although water- CO_2 experiments illuminate the behavior of explosively degassing systems,

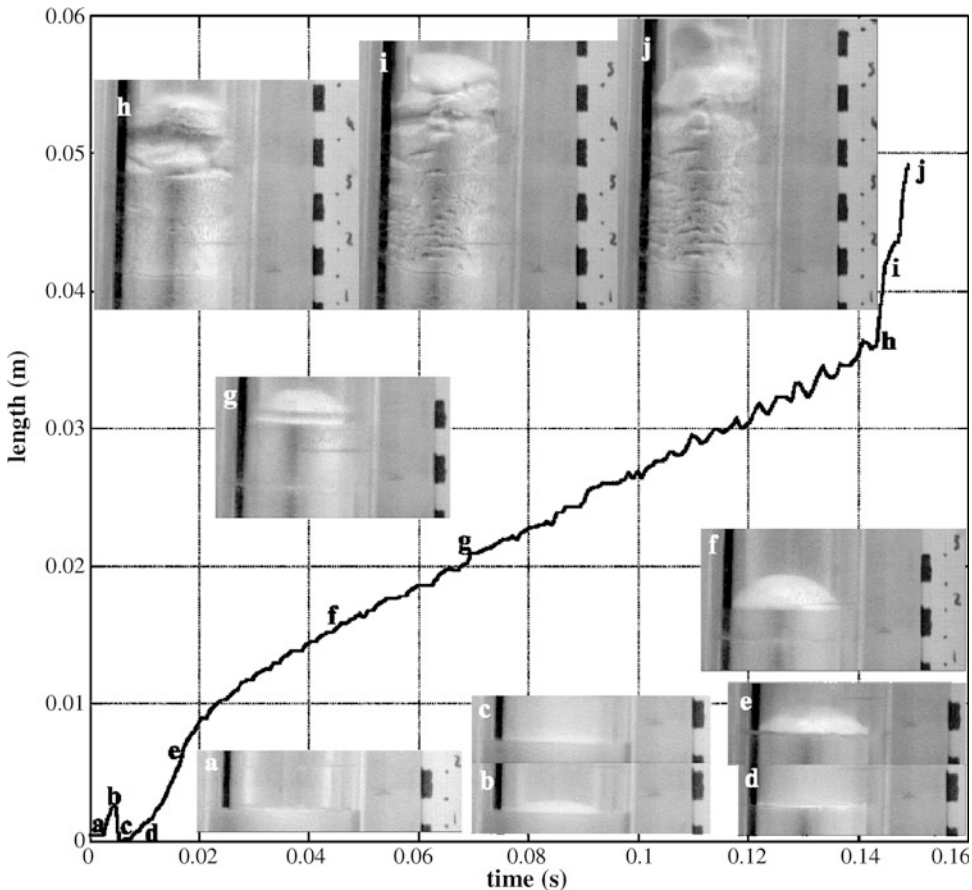
water is many orders of magnitude less viscous than magma. Zhang et al. (1997) investigated the role of viscosity by adding polymers to the water, increasing liquid viscosity by between 1 and 3 orders of magnitude. Degassing styles ranged between passive degassing with little or no expansion for low supersaturation decompressions, through to rapidly expanding and fragmenting flows when pressure reduces by a factor of 50 or more on decompression. Bubbles were found to appear uniformly throughout the liquid on decompression, and were considered to be the result of heterogeneous bubble nucleation on minute particles in the liquid. This differs from boiling liquid flows and is more consistent with heterogeneous bubble nucleation on crystals in magmas. Initial acceleration of the liquid was found to be constant if ambient pressure was constant, with bubbles growing as $\text{time}^{2/3}$. Viscosity was found to be less important than the dissolved gas content in determining explosive capability, indicating that the low solubility of CO_2 in magmas made explosive eruptions driven solely by CO_2 unlikely. Experimental bubble growth was found to match well with models formulated for bubbles in magma, adding to confidence that the experimental system was, at least in part, similar to the natural phenomena.

Taddeucci et al. (2006) increased complexity by carrying out degassing experiments with a viscoelastic analogue material, i.e., one that can change from viscous liquid to brittle solid behavior as strain rate increases. The silicate analogue was polydimethyl siloxane, with various additives, sold commercially as ‘Silly Putty’. Silly Putty is a dilatant material with yield strength, and is viscoelastic (Cambridge Polymer Group 2002). The volatile analogue was the inert gas Argon, which had minimal effect on the rheology of the Silly Putty. Pressures up to 10 MPa were also found to have minor effect on rheology. The apparatus used was similar to the ‘Fragmentation Bomb’ (Fig. 8) used for decompression of non-degassing samples; however, because the Silly Putty experiments were carried out at room temperature, a visibly transparent shock tube could be used to facilitate high-speed imaging. Samples were loaded into the base of the shock tube, then

exposed to Ar gas at pressures varying between 4 and 13 MPa, for times between 5 and 340 hours, then rapidly decompressed to 0.1 MPa.

Figure 22 shows the flow response on a depressurization of 11 MPa after 5.7 hours of exposure to Ar gas. On decompression, a small upward bulging of the sample surface was followed by detachment of the upper part of the sample from

the tube walls. This is indicated by the change in wall refraction between Fig. 22c, d as the Silly Putty ‘unsticks’ from ‘wetting’ the tube wall and allows a thin film of gas to change the refractive index structure of the interface. The detachment possibly represents a solid-like radial contraction to axial extension (Poisson’s ratio) caused by elastic expansion of the bubble-free sample on



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 22 An 11-MPa rapid decompression of viscoelastic ‘Silly Putty’, saturated with Ar gas (Taddeucci et al. 2006), resulted in an expanding degassing flow. The domed flow top indicates the presence of significant wall friction, despite optical indication that the ‘Silly Putty’ did not wet the tube wall during expansion (compare images c and d). Constant velocity expansion then occurred (e to h) with a circumferential fracture forming near the flow front at (g). Flow front oscillation is apparent between g and h, but no concurrent pressure data are reported. From (h), the flow front velocity increases dramatically as pervasive brittle fractures appear in the flow margin (i and j) where strain rates are highest.

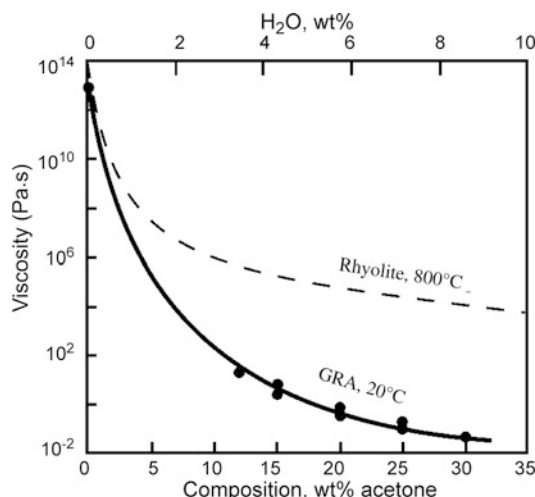
Fragmentation did not occur because the fractures did not propagate through the flow center where strain rates were low and viscous behavior persisted. The fractures could then close and heal due to viscous flow once strain rate declined. These experiments give valuable insight into the process of margin fracture, which could provide an efficient means for gas escape, thus suppressing explosive behavior. They also demonstrate a mechanism of generating fluctuations in magma effusion rate. At volcanic scale, margin fracture may be expected to generate a seismic response, and field evidence of such processes exists (Tuffen and Dingwell 2005). (Reprinted with kind permission from Elsevier)

decompression. Alternatively, the momentary appearance of doming of the sample surface suggests a significant degree of friction with the tube wall and the generation of high strain rate; the Silly Putty wetting of the tube wall breaks down, but wall friction remains significant. Bubbles nucleate and start to expand. The sample surface domes up, suggesting that despite detachment, wall friction is still appreciable (compare with Fig. 20). The sample then expands and develops fractures at the tube wall. Sharp fractures near the flow head were defined as ‘rim fractures’, with other more numerous and widespread fractures called ‘pervasive fractures’. These fractures represent a solid-like response to the high strain rate imposed at the flow margin and give experimental support for a theoretical process proposed by Gonnermann and Manga (2003) who postulated that fragmentation was occurring due to viscous shear along flow edges. The fracturing process may also create seismic signals (Tuffen and Dingwell 2005) and act as a trigger for seismically detectable resonant phenomena. The fractures only propagate as far as strain rates maintain brittle behavior, which acts to confine fractures within the outer section of the flow. Fragmentation will occur if the strain rate is sufficient for brittle behavior to persist right to the center of the flow. The fractures were also observed to close and heal because, once formed, lower strain rates caused reversion to viscous behavior. The development of such fracture networks has significant implication for mechanisms of magma degassing. The continual formation and healing of margin fractures gives rapid access to the conduit margin for gas deeper in the flow. This has consequences for transitions between effusive and explosive behavior.

Only a few Silly Putty experiments underwent explosive fragmentation. All these were triggered at a sudden tube widening, emphasizing the importance of conduit geometry on flow behavior in volcanic systems. In comparison, similar decompressions of pre-vesiculated samples (Ichihara et al. 2002) resulted in much more extensive fragmentation. This demonstrates that, although there may be more gas dissolved in an unvesiculated sample than contained in the pores

of another non-degassing sample, the pre-vesiculated material is more explosive. The process of bubble nucleation and diffusive growth provides a regulating mechanism on the effective maximum rate of decompression and reduces initial explosive activity. However, the continued supply of volatile into bubbles means that degassing flows are likely to expand more given sufficient time, but in the absence of geometric structures, will they fragment in a brittle solid manner?

The viscosity of the liquid phase in magma increases by several orders of magnitude (Fig. 1) as water degasses, also decreasing the strain-rates over which viscous flow turns to brittle solid behavior. Water and melt interact to change the lengths of silicate molecules and ‘lubricate’ differential molecular motion, hence changing viscosity. As an analogue material, gum rosin derives from removing volatile compounds from tree sap, leaving 3-ring aromatic hydrocarbons typified by abietic acid ($C_{19}H_{29}COOH$). It is a glassy organic solid at room temperature, but melts in the region of 80 °C, and becomes increasingly less viscous as temperature rises. Warm gum rosin behaves in a viscoelastic fashion (Bagdassarov and Pinkerton 2004), being brittle when deformed rapidly but viscous under more gentle treatment. Such behavior is phenomenologically similar to a silicate melt. Phillips et al. (1995) found that certain oxygen-bearing organic solvents, such as acetone, dissolved in gum rosin to yield a liquid at room temperature. At concentrations below about 20% w/w solvent these appear to be true solutions (Lane et al. 2008), with a partial pressure of acetone below that of the pure liquid suggesting some chemical interaction; another phenomenological similarity to hydrated magma. The viscosity of this liquid was found to be a non-linear function of the acetone content of the solution, again similar to a hydrated silicate melt (Fig. 23). Decompression of gum rosin-acetone solution below the saturated vapor pressure of acetone (about 20 kPa at room temperature) was found to result in volatile degassing and consequent increase in liquid viscosity. Slow decompression resulted in the establishment of a degassing interface that consumed the liquid solution. No measures were



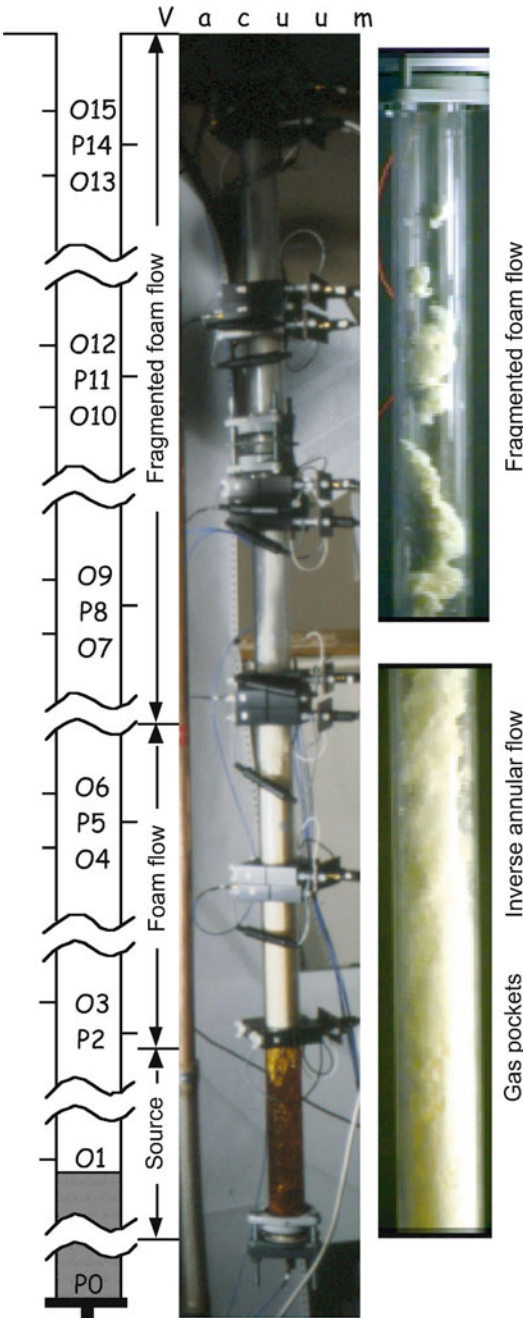
Volcanic Eruptions, Explosive: Experimental Insights, Fig. 23 The viscosity of hydrated silicate melts changes non-linearly with water content (Fig. 1). One experimental system uses gum rosin as an analogy for silicate melt, and organic solvents such as acetone and diethyl ether instead of water. The degassing of gum rosin acetone (GRA) results in increasing liquid viscosity (lower scale), in similarity with the degassing of hydrated magma (upper scale) (Mourtada-Bonnefoi and Mader 2004). The volatile content between natural and analogue systems can be compared by quantifying in terms of moles of volatile per unit volume of liquid; 4% water w/w in magma equates to about 20% w/w acetone in gum rosin. The scales here are not matched on this basis. (Reprinted with kind permission from Elsevier)

taken to remove nucleation sites from the liquid, suggesting that nucleation was sensitive to pressure and was confined to a surface layer in the liquid. This is similar behavior to boiling experiments (Hill and Sturtevant 1990; Sugioka and Bursik 1995), where the reaction force from expansion of degassing volatile and accompanying liquid was considered to elevate pressure in the underlying liquid and suppress bubble nucleation and growth, therefore acting to confine rapid degassing to a thin interface. This process may not be as prevalent in H₂O-CO₂ and Silly Putty systems because absolute pressures were larger, in relation to hydrostatic, than in gum rosin-acetone or boiling refrigerant experiments.

Rapid decompressions of gum rosin-acetone resulted in expanding foam growing from the liquid surface (Phillips et al. 1995). Flow-front acceleration was generally constant, except for

large supersaturations with high acetone content where accelerations increased with time. These were reinterpreted by Mourtada-Bonnefoi and Mader (2001) in the light of additional experiments that found bubble growth prior to fragmentation is diffusion-driven, but limited by the high-viscosity skins that develop around growing bubbles because of the reducing volatile concentration. This infers that diffusion-limited degassing is unlikely before fragmentation removes viscous control as a physical mechanism. Accelerations and velocities using gum rosin-acetone were an order of magnitude less than those produced by CO₂-H₂O experiments, but this is consistent with the order of magnitude lower driving pressures. Experiments carried out with high supersaturations allowed extensive acetone degassing to take place. This resulted in the production of solid gum rosin foam with elongated bubbles, phenomenologically similar to woody or tube pumice.

The experiments of Phillips et al. (1995) were carried out using small volumes of solution in un-instrumented and relatively short tubes that did not allow sufficient time or length for flow patterns to develop and stabilize. Lane et al. (2001) carried out similar experiments using longer tubes, larger volumes of gum rosin-acetone and gum rosin-diethyl ether solutions, and high-speed pressure measurement. Instantaneous decompression from 10⁵ to 10² Pa resulted in rapid degassing from the liquid surface. Flow became pseudo-steady after about 0.2 s, with foam expanding from the liquid interface, and continued explosive behavior for about 1.5 s. Pockets of gas developed at the tube wall, eventually merging to allow the expanding foam to detach from the wall. Removal of wall drag allowed the foam core to accelerate as an inverse annular, or detached flow, and break up into foam fragments that were ejected from the top of the shock tube (Fig. 24). The fragmentation of the foam core was precluded during expansion of a non-degassing system (Namiki and Manga 2005), suggesting that degassing systems have different flow patterns to those driven by gas expansion alone. The reason for this difference may lie with the smaller overpressure in bubbles of



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 24 An 0.1-MPa rapid decompression of a solution of gum rosin 19% w/w diethyl ether generates a degassing flow (Lane et al. 2001). Bubble growth occurs predominantly at the upper liquid surface. The expanding flow develops gas pockets at the wall, which merge and allow a central foam core to detach from the tube wall giving inverse annular flow. The detached foam core is no longer constrained by wall drag and expands until it

pseudo-steady degassing systems compared with those in transient expanding flows. The lower bubble pressures prevent bubble wall breakage and stabilize detached flow until fragmentation commences (Namiki and Manga 2005). This suggests that flow behavior preceding fragmentation may differ between explosive volcanism triggered by rapid decompression of vesiculated magma, for example the May 18, 1980 Mt. St. Helens lateral blast triggered by a landslide, and sustained explosive volcanism such as the Plinian eruption which followed the blast. The fragmentation mechanism may also be different, with brittle spalling fragmentation before expansion (Figs. 9, 16 and 17) suggested for the lateral blast and fluid expansion before ‘plastic’ fragmentation (Figs. 19, 20 and 24) for the Plinian eruption.

The size of fragments of gum rosin foam was initially significantly smaller than tube diameter, but increased to near tube diameter and elongated with time, suggesting a declining flow energy. Interestingly, some of the flow features found by Cagnoli et al. (2002) in expanding particle beds appear similar to those in expanding gum rosin foams, with bubbles appearing at the margins of both flows and being stretched out (Fig. 15). The sequence of flow pattern changes in the gum rosin flows was found to result in the generation of characteristic pressure fluctuations. Pressure fluctuations may be inherent in the unsteady nature of flows undergoing changes in behavior and such phenomena could be detectable by seismic means during volcanic flow.

Magma can fragment in the presence of pre-existing bubbles alone without any degassing (e.g., Fig. 9), and the presence of crystals increases the number density of nucleation sites (e.g., Fig. 4), and hence the surface area over which degassing can occur. Magmas are decompressed with crystals and bubbles often present, an aspect investigated analogously by

←
Volcanic Eruptions, Explosive: Experimental Insights, Fig. 24 (continued) breaks up to give a fragmented flow. The fragments of gum rosin foam are ejected from the shock tube. (Reprinted with kind permission of AGU)

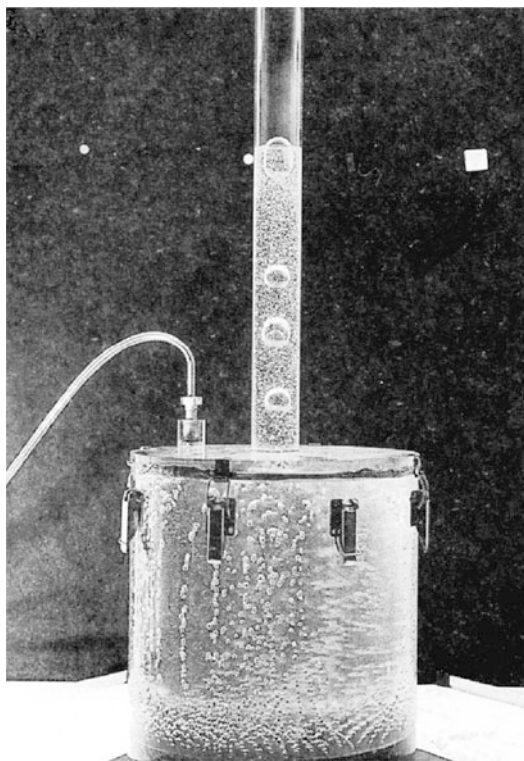
Mourtada-Bonnefoi and Mader (2004) using gum rosin-acetone mixtures and a variety of particles and existing bubbles. It was found that fragmentation occurred at lower volatile contents and lower decompressions in the presence of particles and bubbles. It was also found that flow behavior depends on the distribution of particles in the liquid, with more rapid degassing where particle concentrations were high. This suggests that heterogeneity in the distribution of crystals and pre-existing bubbles will develop into sizable features within a degassing flow, at least before fragmentation. This raises the possibility that flow heterogeneity over a range of size scales may then play a role in the fragmentation process.

Low VEI Events High VEI events do not occur very frequently compared to anthropogenic timescales, and do not lend themselves readily to detailed field investigation of subterranean flow processes. Conversely, low VEI events, typified by Strombolian activity, occur on anthropogenically short timescales and form relatively safe natural laboratories to study magma flow in the volcanic conduit. Low VEI eruptions commonly occur at volcanoes where magma viscosity is low and degassing water is able to separate relatively easily from the magma; indeed, this separation is required to produce low VEI events. Analogue experiments of separating flows may, therefore, be compared with relatively well-studied natural processes in ways that high-VEI events cannot. Experimental conditions are also simplified in relation to those applied to high-VEI processes, with the assumptions that liquid rheology and temperature remain constant being easier to justify. Low-VEI processes may also be studied without shock tube or high-pressure apparatus, making observation and measurement more straightforward.

Strombolian eruptions are characterized by short-lived explosive events that erupt significantly higher water to silicate mass ratio than that of the bulk magma (Blackburn et al. 1976; Chouet et al. 1974). These events are separated by longer periods of relative quiescence, suggesting the presence of a gas separation process that

produces gas-rich regions that rise up the conduit between intervals of relatively gas-depleted magma. In two-phase flow terminology, one such regime is that of slug flow, where large ascending gas slugs can form from the coalescence of many much smaller bubbles. However, evidence from extensive research on water-gas systems indicates that for such a flow pattern to be stable in the conduit the overall gas volume fraction would probably need to exceed about 0.25 (Clift et al. 1978), and the time intervals between explosions would be similar to, or less than explosion duration. It is also very questionable that slug flow would form from numerous small bubbles in a relatively high viscosity liquid like basalt magma. This suggests that other mechanisms operate to promote formation of gas slugs. Jaupart and Vergnolle (1988, 1989) were among the first to recognize that volcanic conduits are not straight vertical tubes, but exhibit significant geometrical heterogeneity that may play a major role in eruptive behavior. Experiments were carried out to examine the effect of a flat roof in trapping the numerous small bubbles rising through the magma (Fig. 25). Small air bubbles were injected at variable rate into the base of a tank containing liquids of various viscosities. The bubbles rose to the roof of the tank, which had a vertical small diameter outlet tube. Three modes of system behavior were observed. At lower liquid viscosity, a raft of bubbles formed at the roof. At some critical thickness the bubble raft collapsed on the scale of the tank roof and flowed up the outlet tube as a large gas bubble. This behavior was considered analogous to Hawaiian eruptions. At higher liquid viscosities, the bubble raft collapsed on a scale smaller than the tank roof and a series of bubbles that are much larger than those injected, but smaller than the low viscosity case, emerge up the outlet pipe. This gives insight into Strombolian eruptive activity. The final mode of behavior was the development of a relatively stable bubble raft and the escape of small bubbles up the outlet pipe to give bubbly flow at the surface.

The ascent of large gas bubbles, or slugs, up a tube can be controlled by surface tension, liquid viscosity or liquid inertia. In basaltic systems



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 25 Numerous small nitrogen bubbles rising in a silicone oil of viscosity 1 Pa s (Jaupart and Vergnolle 1988) encounter a structural discontinuity in the form of the horizontal roof of the experimental tank. Bubbles trapped under the roof collide and coalesce to emerge up the narrow exit pipe as much larger bubbles. Such a mechanism can be envisaged for the production of large bubbles some 1–3 km beneath the vent at Stromboli volcano (Burton et al. 2007); however, the experimental geometry canonically demonstrates the process rather than simulating conduit structure. Large bubbles emerging from the trap then expand and become overpressured on approach to the surface (James et al. 2008). (Reproduced with kind permission of Cambridge University Press)

surface tension has negligible control, and inertia dominates with some component of viscous influence (Seyfried and Freundt 2000). Conveniently, this allows the use of water and air at laboratory scale to give inertia-controlled slug flow. Seyfried and Freundt (2000) carried out experiments applying the flow patterns that develop with increasing air flux through a column of water to basaltic volcanism. Single gas slug ascent was found to only occur through sudden incidents of limited volume gas release, with one mechanism

being that of foam collapse from a trap (Jaupart and Vergnolle 1988, 1989). It was observed that slug burst in the tube was a gentle process, but that if the liquid level was in the reservoir atop the tube (akin to a lava lake) then burst was more vigorous and could be likened to Strombolian activity; another example of the impact of conduit geometry on flow. Experiments with gas traps in the water column found that the coalescence of trapped gas pockets resulted in significant pressure oscillations that were a potential source mechanism for volcanic tremor; linking volcano-seismic signals to fluid flow could be a powerful method of experimental scaling inaccessible in high VEI events.

Seismic signals may result from the collapse of a foam layer trapped at a roof. Measurements made within apparatus similar to that of Jaupart and Vergnolle (1988, 1989) reveal pressure fluctuations both within the fluid and in the air above the fluid surface in the outlet tube (Ripepe et al. 2001). These could be linked to seismic and acoustic signals measured at Stromboli volcano and give insight into the role of fluid-flow on the generation of seismo-acoustic waves during explosive volcanism. During the build up of the foam layer, no pressure changes were detected. Collapse of the foam layer and motion of the gas slug up the outlet pipe produced pressure changes in the liquid tank of a few-hundred Pa, and of a few-tenths Pa in the air above the liquid level. Collapse of the foam layer generates upward gas flow in the outlet pipe thereby producing downward motion of the dense liquid phase in the outlet pipe, reducing pressure. Oscillation of liquid on the gas spring of the slug produces pressure oscillations within and above the liquid. Slug burst results in a sudden increase in oscillation frequency above the liquid, and a small pressure pulse in the liquid due to decay of liquid flow around the slug nose (James et al. 2004). Ripepe et al. (2001) postulate that seismic signals measured at Stromboli are generated during rapid gas expansion during foam collapse, and estimated that the slug be about 4.5 m long and induce a pressure drop of 0.1 MPa in the magma.

Inversion of highly repeatable seismic waveforms measured at Stromboli Volcano, Italy

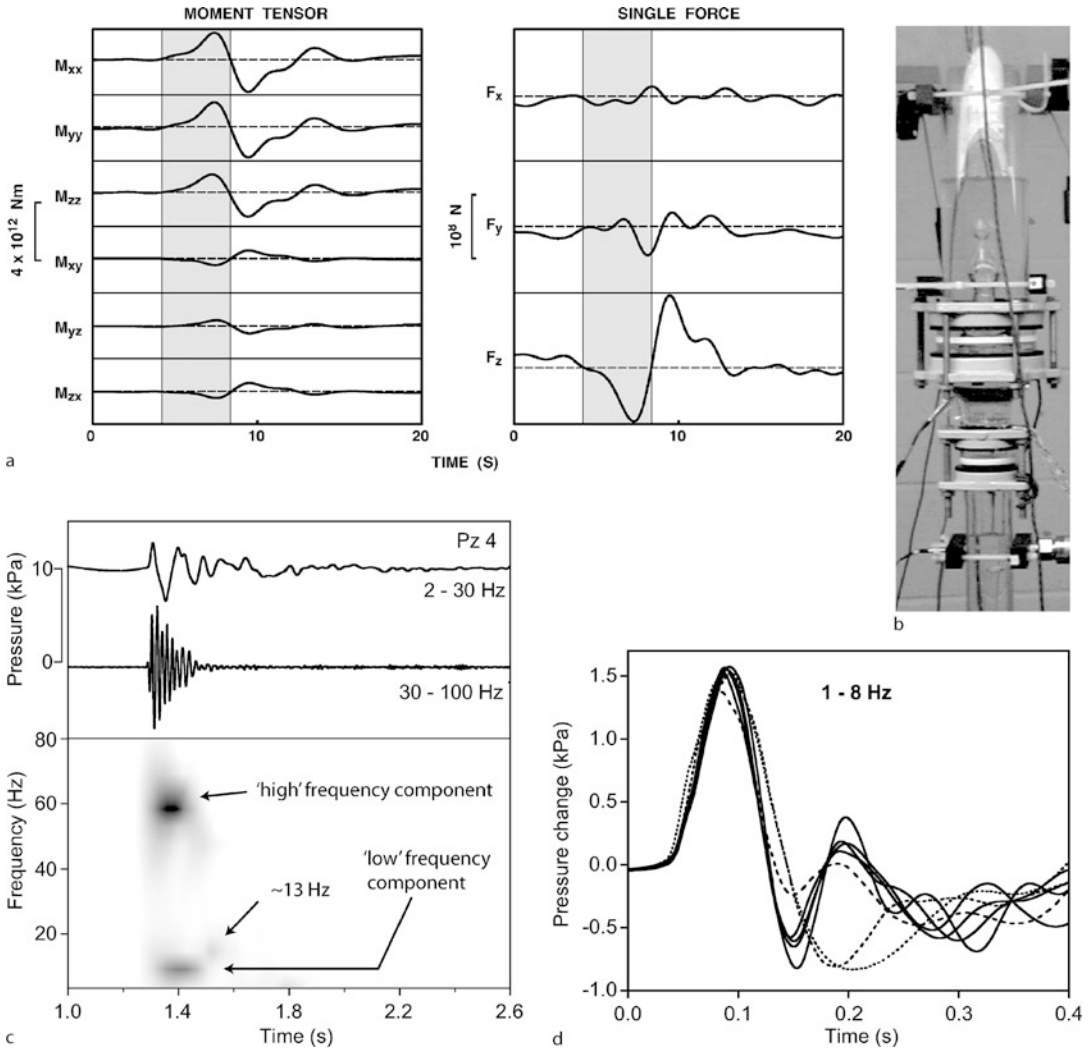
(Chouet et al. 2003, 2008) has enabled imaging of the conduit as well as revealing the forces and pressures being created by magma flow during Strombolian eruptions. In summary, Stromboli's plumbing system appears to comprise a dyke dipping about 72° that runs from the vent system to an intersection with a dyke dipping about 45° and 900 m below the vent system. Stable and repeatable seismic sources are inferred at a depth of about 220 m below the vent system, and have the characteristics of a dyke segment intersecting the main conduit. Another, weaker, seismic source is located at the intersection of the two main dykes about 900 m below the vents. The dominant seismic source at 220 m below vent generates a downward vertical force of about 10^8 N followed by an upward force of similar magnitude (Fig. 26a). The source volume increases during the downward force and decreases with the upward force. These measurements provided a means of testing laboratory models of Strombolian eruptions.

James et al. (2004) studied pressure oscillations resulting from the ascent of single gas slugs in a vertical tube. The oscillations observed at the top of the liquid column (Fig. 27) showed an increasing amplitude and decreasing period as the slug ascended, with oscillations ceasing as the slug reached the surface. Burst of the meniscus produced further pressure fluctuation. Modeling of the oscillations revealed the source mechanism as bouncing of the liquid mass above the slug on the spring of gas within the slug. The observed pressure oscillations were not similar to the change in volume of the VLP seismic source at Stromboli (Fig. 26a), nor were oscillation amplitudes sufficient to account for forces of 10^8 N at volcanic scale. Bubble bursting was also quiescent rather than explosive.

Such experiments on Strombolian processes were not scaled for expansion of the gas slug as it rises, and gas chemistry suggests that slug source depths could be in the range 900–2700 m (Burton et al. 2007). A slug ascending from 1000 m in magma will expand about 200 times to reach atmospheric pressure, but in the experiments expansions were in the region of 1.2. In order to investigate more appropriate values,

James et al. (2008) used low vapor-pressure vacuum pump oil as an analogue liquid and varied the ambient pressure above the liquid surface between 10^5 and 10 Pa. The reduction in ambient pressure provided a proxy for slug ascent from depth and allowed potential gas expansions of factors up to 10^4 . The positions of the slug base, slug nose and liquid surface for ascending slugs of various gas masses and ambient pressures are shown in Fig. 28. Regardless of ambient pressure and slug mass, the ascent velocity of the slug base remained essentially constant. However, the slug nose accelerated rapidly on approach to the surface when ambient pressure was low, and burst with a pressure higher than ambient. Gas slugs that underwent little expansion burst at near ambient pressure. Scaling to the volcanic case was undertaken using experimentally verified CFD modeling. This showed that slug burst pressure is a function of conduit radius and slug mass, with large slugs in narrow conduits producing gas overpressures in the region of 1 MPa at burst. This overpressure will fragment the upward-moving (tens of m s^{-1}) meniscus of liquid existing just prior to burst. The effect of such decompression on the vesiculated basalt magma flowing down the conduit walls and slug base could be explosive fragmentation (Namiki and Manga 2006). This would eject pyroclasts, adding greatly to the pyroclast volume from the meniscus and explaining how explosive slug burst results in the ejection of significant masses of pyroclasts. There is also the possibility that the gas slug is not a single bubble, but more accurately described as a foam raft, although such a structure may not survive the degrees of expansion undergone in ascent from 1000 m depth with its bubble walls intact. The rapid 1 MPa decompression of a foam raft would also produce pyroclasts as the foam fragments.

Rapid slug expansion, which generates the explosivity of Strombolian eruptions, is a process operating within the top few tens of meters of the magma column (Fig. 28c). This, and the fact that the forces generated are of order 10^6 N upward, means that slug expansion is not the fluid dynamic source process responsible for the VLP seismic signals inverted by Chouet et al. (2003). Slug

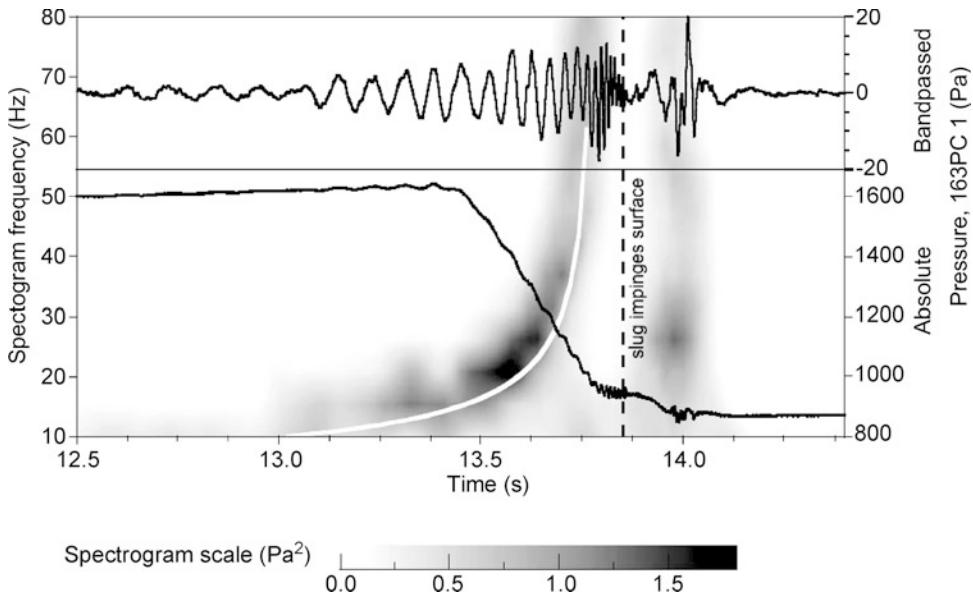


Volcanic Eruptions, Explosive: Experimental Insights, Fig. 26 The geometry of experimental tubes has been observed to change flow behavior in a number of instances. James et al. (2006) investigated the dynamics of a gas slug ascending through a tube widening. Inversion of seismic data measured at Stromboli volcano (a, Chouet et al. 2003) indicated an initial downward force of about 10^8 N, accompanied by pressure increase expanding the conduit. This was followed by upward force and conduit contraction with a period of about 10 s. The experimental ascent of a gas slug through a tube widening (b) caused breakup of the slug and production of a transient gas jet.

Spectral analysis of pressure (c) showed three resonant components. The low frequency component (d, about 6 Hz) showed one to two cycles of pressure increase followed by decrease, qualitatively consistent with pressure change in the conduit at Stromboli during explosive eruption. Measurement of apparatus displacement indicated a peak downward force of 30 N, which when naïvely scaled by 100^3 for physical dimension and 2.5 for density difference gives a force of 7.5×10^7 N, encouragingly close to that estimated from seismic inversion. (Reprinted with kind permission of AGU)

expansion probably plays little role in this particular seismic signal because the magma head is probably too large for the slug to be expanding as any appreciable rate. The stable position of the

seismic source suggests geometric control of a flow process, a phenomena we have seen a number of times before. James et al. (2006) measured pressure and force as a gas slug ascended from a



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 27 Strombolian eruptions are short-lived, periodic and have high volatile content. For these reasons they are associated with the formation, ascent and bursting of large bubbles of water vapor (often known as slugs), rather than the expansion of a vesicular magma (Fig. 21). The ascent of air slugs in water during small-scale experiments caused pressure oscillations (James et al. 2004).

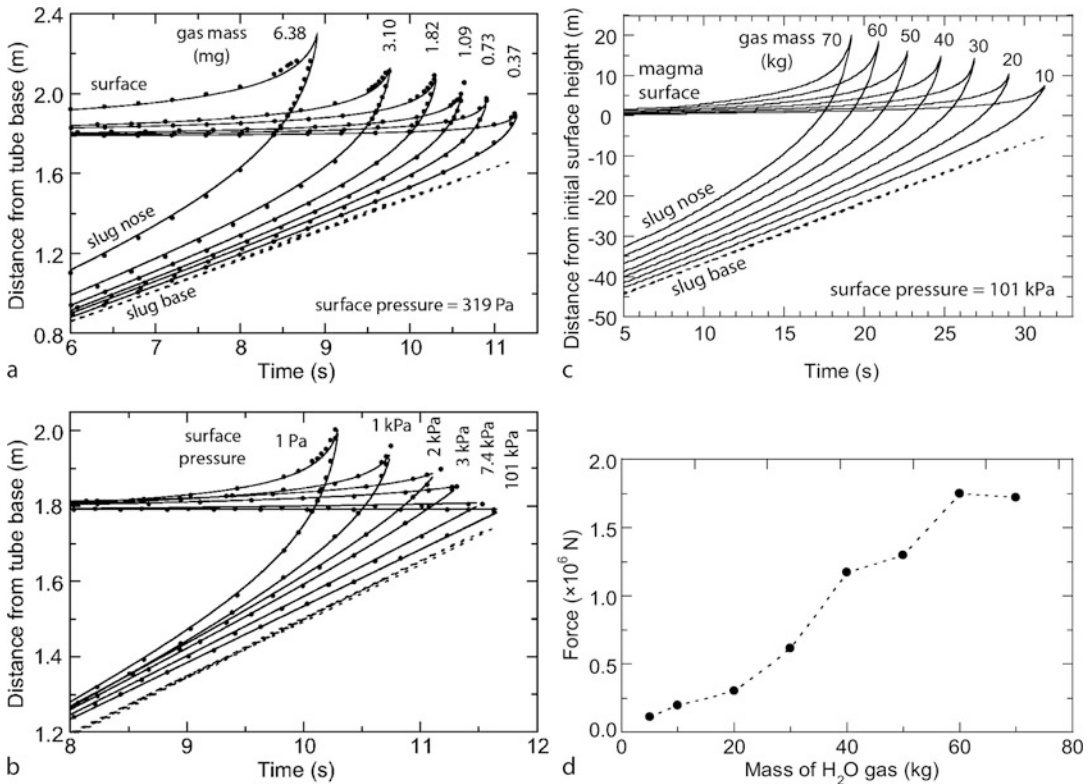
A spectrogram (gray shading) of band-passed (upper trace) pressure data (lower trace) was found to match well with calculations of the oscillation frequency (white trace) of the water above the gas slug bouncing resonantly on the gas slug. This oscillatory mechanism could not be used to account for the seismic signals used to image the conduit at Stromboli (Chouet et al. 2003). (Reprinted with kind permission from Elsevier)

narrow tube through a flare into a wider tube. The gas slug segmented on passing through the flare and liquid flow during this process generated pressure oscillations (Fig. 29). Analysis of these oscillations (Fig. 26) revealed three components centered on 60, 13 and 6 Hz, with the low frequency component showing increasing pressure followed by a decrease, resembling the expansion-contraction waveform from inversion of VLP seismic data (Fig. 26a; Chouet et al. 2003). The vertical force generated during the experimental pressure increase was 30 N downward. Applying a simplistic scaling based on density and volume differences between laboratory and volcano gives a force in the region of 10^8 N, of similar magnitude to that emerging from inversion of VLP seismic data. It was, therefore, suggested that the source mechanism for VLP seismic signals measured at Stromboli by Chouet et al. (2003) was the ascent of slugs

sourced at depth (Burton et al. 2007) through a geometric widening of the feeder dyke inclined at 72° . The upward widening disrupts the downward flow of liquid around the ascending slug, allowing significant thickening of the falling film. As this thickening falls through the downward narrowing, it pinches closed and segments the slug. The resulting liquid piston continues to descend because of its inertia, and increases pressure below it. The piston then oscillates for a cycle or two before being disrupted by the buoyancy of the gas beneath. Experimental insight suggests that the source mechanism for the VLP seismic signal is the deceleration of the liquid piston.

Explosion Consequences

Various consequences of explosive degassing of magma have been studied by analogue means, and many of these are reviewed in Sparks et al. (1997).



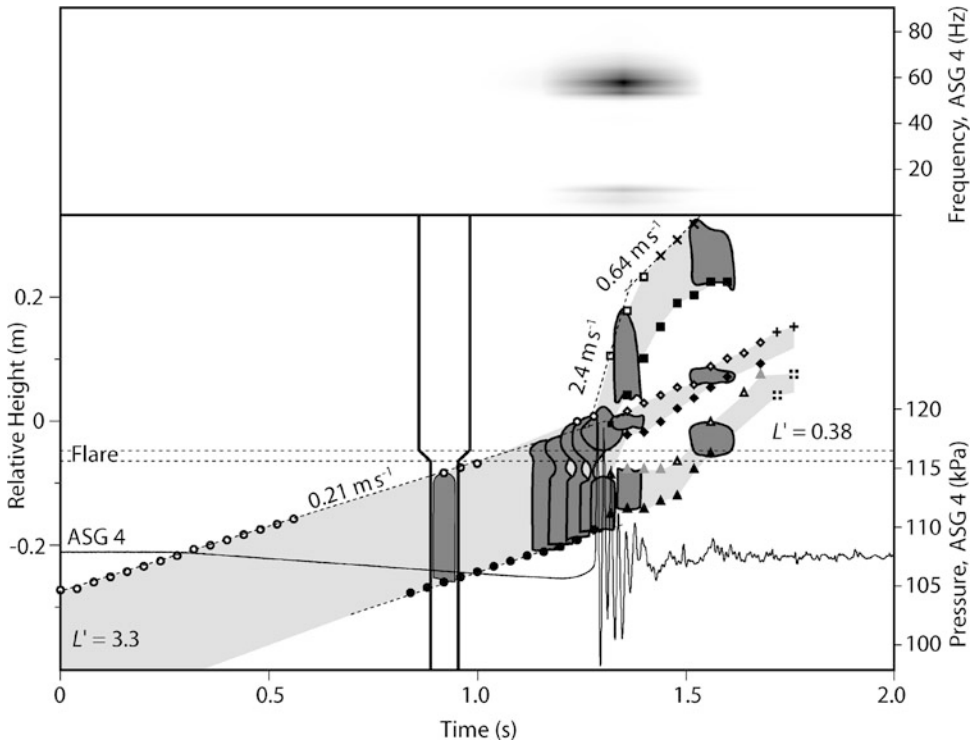
Volcanic Eruptions, Explosive: Experimental Insights, Fig. 28 Experiments on slug flow are numerous in the engineering literature, but volcanic flows operate under conditions not investigated for anthropogenic purposes. One of these conditions is the large pressure change a gas bubble will experience in rising from some depth. James et al. (2008) investigated the dynamics of rapid gas slug expansion as a result of this decompression using an air bubble and vacuum pump oil as analogue fluids. Using surface pressures below atmospheric allows the air bubble to undergo considerable decompression under laboratory

conditions. The degree of slug expansion, and hence explosivity on burst, was found to depend on (a) the mass of gas in the slug as well as (b) the applied surface pressure. Computational fluid dynamic modeling was used to apply the experimental findings at volcanic scale (c). The process of slug expansion generated significant pressure change and vertical force (d), but the amplitudes are two orders of magnitude less than those calculated at Stromboli (Chouet et al. 2003, 2008). (Reproduced with kind permission of the Geological Society of London)

Here, we review a sample of more recent literature.

Cinder Cones Explosive volcanic eruptions produce pyroclastic constructs, the most common form being known as cinder or scoria cones. The basis for understanding cinder cone growth was given by McGetchin et al. (1974). These structures, which are piles of generally unconsolidated magma fragments, are prone to gravitational collapse and rapid erosion. To test the hypothesis that cinder cone geometry is controlled by the behavior of unconsolidated fragments, Riedel et al.

(2003) carried out small-scale analogue experiments using grains of sugar and sand. These were built into cones, drained in the center to represent the crater. These grain piles showed similar geometry to natural cinder cones (Fig. 30). Quantification of the geometry showed that analogue and natural cones were identical in shape to within 5%. This demonstrates that the angles of repose exert fundamental control on cinder cone geometry, with surface grain flow maintaining angle of repose during cone growth. The angle of repose nature of cinder cones emphasizes their vulnerability to erosion and collapse.



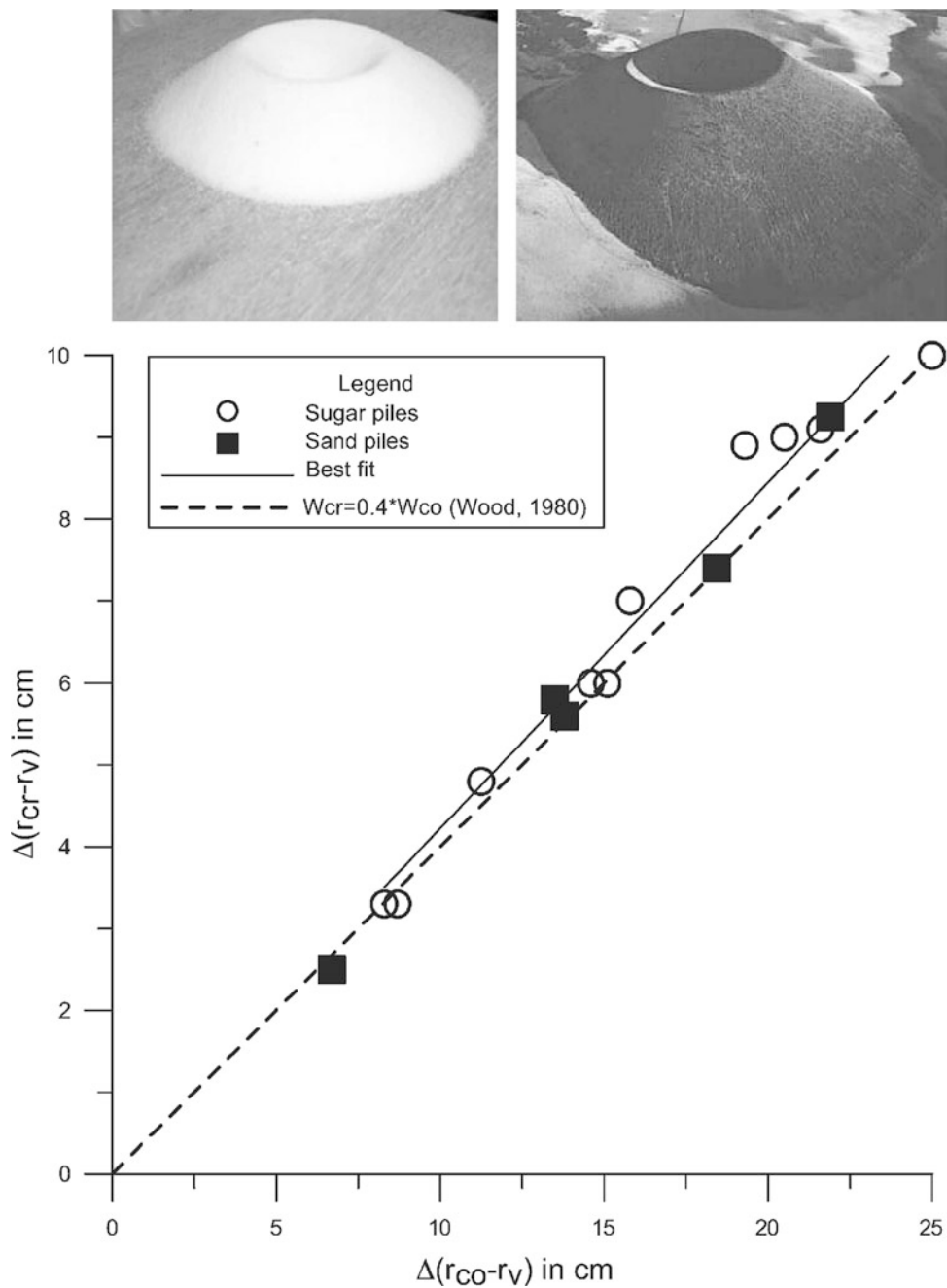
Volcanic Eruptions, Explosive: Experimental Insights, Fig. 29 The source mechanism suggested by laboratory experiment (Fig. 26) for seismic signals measured by Chouet et al. (2003) at Stromboli Volcano, Italy is the creation and rapid deceleration of a downward-moving liquid piston, formed as a gas slug breaks up in a conduit widening (James et al. 2006). The liquid piston ‘bounces’ on the lower portion of the slug for a cycle or two before

being consumed by the ascending remains of the original slug. The cartoon superimposed on the pressure trace correlates the position of gas bubbles with pressure measurement and spectral analysis of pressure fluctuations. The dramatic velocity increase of the slug nose does not appear to play an obvious role as a source mechanism and may be associated with the highly unstable bubble in the tube widening. (Reprinted with kind permission of AGU)

Pyroclastic Flows Pyroclastic flows ‘hug’ the ground because their density is greater than that of the atmosphere. Gravity then drives the pyroclastic material in a lateral fashion; hence pyroclastic flows are part of the family of gravity currents, and the sub-family of particle-driven flows that includes turbidity currents. The flow behavior of pyroclastic flows, which result from explosive volcanism or dome collapse, has been investigated at laboratory scales using analogue materials for a number of years (Sparks et al. 1997). Gladstone et al. (2004) experimentally examined the dynamics of gravity currents that were initially stratified in density using a 3 m long flume. A range of brine solutions of different density were layered into a lock at one end of the flume and released into tap water. The solutions

were used singly, in two or three layers to investigate how stratification can develop or be homogenized in pyroclastic flows and turbidity currents. Flows were visualized by coloring the solutions allowing detection of mixing and stratification.

Flow behavior was found to depend on two dimensionless parameters. Firstly, the relative values of buoyancy between brine solutions, B^* , where most salt is in the lower layer for $0 < B^* < 0.5$, and in the upper layer for $0.5 < B^* < 1$. Secondly, the density ratio of the solutions, ρ^* , which is small for strong stratification, approaches 1 for weak stratification and equals 1 for unstratified conditions. When the lower layer has a greater proportion of the driving buoyancy ($B^* < 0.5$), it can run ahead leading to



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 30 Cinder cones are the proximal expression of fallout from explosive volcanic eruptions. Riedel et al. (2003) found that small piles of sugar (left image) or sand look similar in shape to much large piles of cinder (right image). Comparison of characteristic dimensions of laboratory cones (r_{cr} is distance from central hole to crater

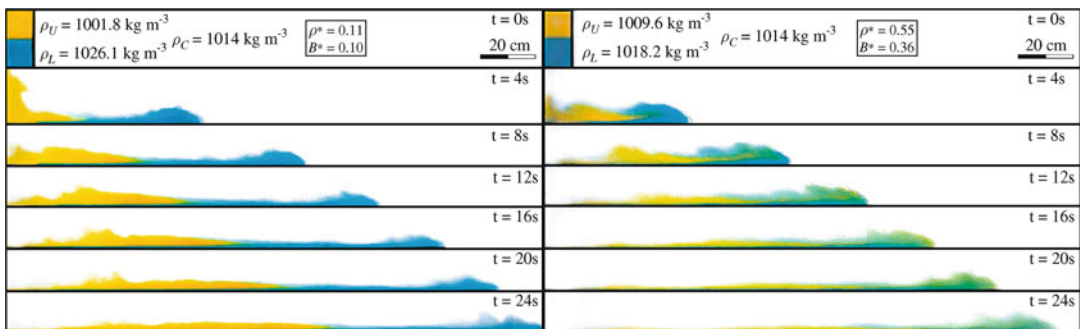
rim, r_{co} is distance from central hole to pile extremity and r_v is vent radius) with established measurements of natural cinder cones confirms the similarity. These experiments showed that cinder cones grow by the surface transport of grains under gravity to leave the slopes at angle of repose. (Reprinted with kind permission from Elsevier)

flow stratification (ρ^* approaches 0, Fig. 31), or the layers can mix to produce a homogeneous current (ρ^* approaches 1, Fig. 31). If the upper layer is more buoyant ($B^* > 0.5$), it travels to the nose of the current by mixing to produce a homogeneous flow (ρ^* approaches 1) or overtaking, leading to flow stratification (ρ^* approaches 0). This suggests that buoyancy-contrast controls separation, whilst density-contrast controls mixing. These observations give insights into the origin of stratification and grain size contrast in pyroclastic flow deposits and allow constraint of flow conditions from field observation.

Caldera Collapse The highest VEI eruptions result from the ejection of large volumes of magma. As magma emerges, country rocks overlying the magma storage area subside, resulting in the formation of a caldera. The caldera roof may fail in a number of modes. Roche and Druitt (2001) carried out a force balance calculation for a caldera roof to fail as a coherent single piston, predicting that $\Delta P/\tau_c \geq 4R$, where ΔP is the pressure below lithostatic at which failure occurs, τ_c is the rupture shear stress and R is the ratio of roof thickness to diameter. Experiments were carried out using sand or a slightly cohesive sand-flour mixture as an analogue to country rock, and a

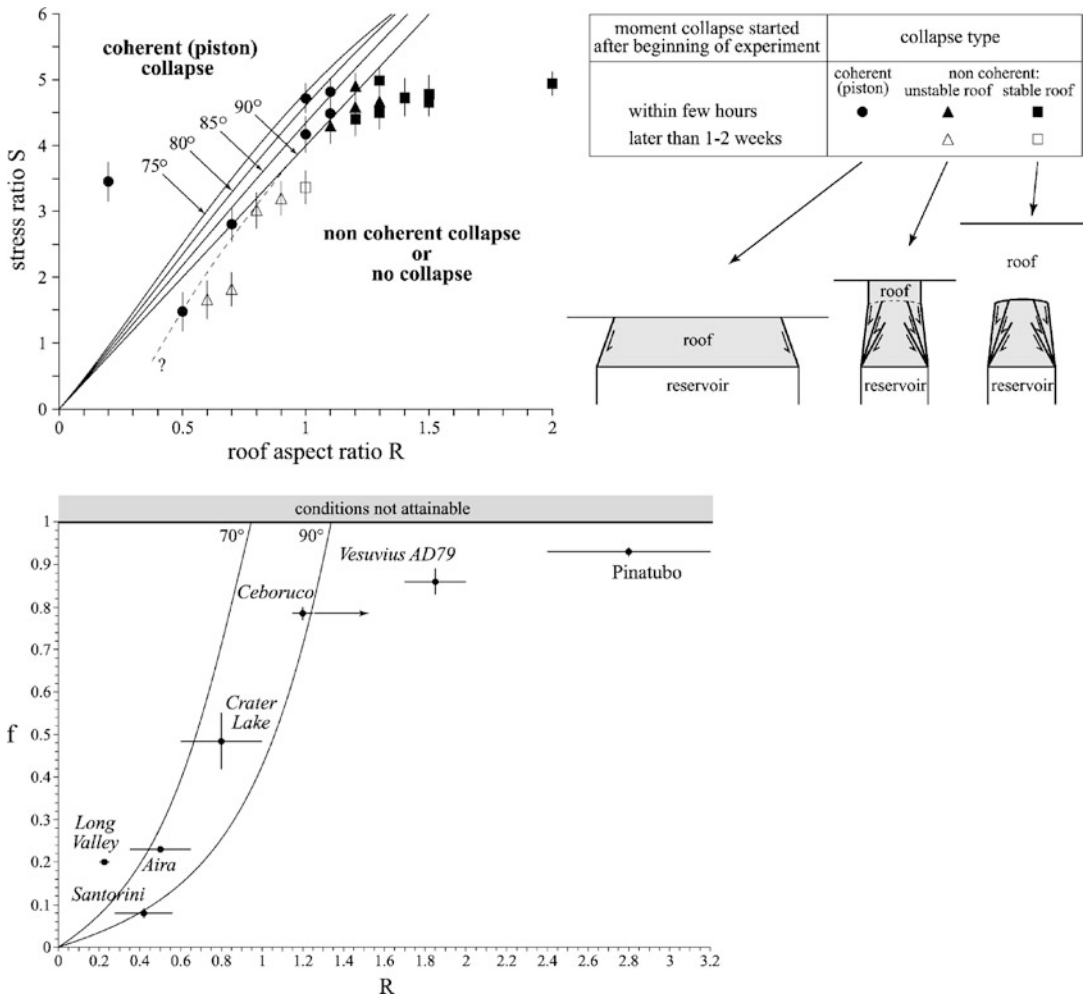
draining reservoir of silicone fluid to simulate removal of magma. Comparison of experiment and theory (Fig. 32) shows that coherent roof collapse did occur in line with theoretical prediction. The experiments also showed two other more complex modes of collapse that occur at roof aspect ratios larger than the coherent collapse field, and that these collapse mechanisms may take considerable time to complete. Comparison of the experimental data with known parameters for seven calderas suggested that four had collapsed as coherent pistons, with the other collapses being non-coherent.

Experiments have helped to constrain possible volumes of caldera forming eruptions from pre-caldera conditions (Geyer et al. 2006), showing a potentially powerful route of predicting the time and size of caldera-forming eruptions. Experiments were carried out using a bed of quartz sand as a rock analogue and a deflating water-filled latex balloon to represent the magma body. Extensive arguments are presented to justify application of the experimental results at volcanic scales. Coherent piston collapse was observed at low roof aspect ratios (roof thickness/diameter), becoming more non-coherent as aspect ratio increased. These experiments showed different



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 31 The first order behavior of gravity currents, which include pyroclastic flows, can be studied by analogue experiment using dyed brine solutions of different density propagating in a water-filled flume tank (Gladstone et al. 2004). Density stratification in the flow source material may be preserved under certain conditions of buoyancy and density (left panel). Pyroclastic deposits from such flows are likely to show vertical and horizontal stratification. Under other buoyancy and density

conditions (right panel) the initial stratification is removed by turbulent mixing and resultant pyroclastic flow deposits are expected to be unstratified. Experiments that reduce complexity, as here by not including suspended particles, often give greater insight into primary controlling parameters. Complexity can then be added to understand the smaller scale processes that lead to diagnostic features identifiable in volcanic deposits. (Reprinted with kind permission from Blackwell Publishing Ltd)

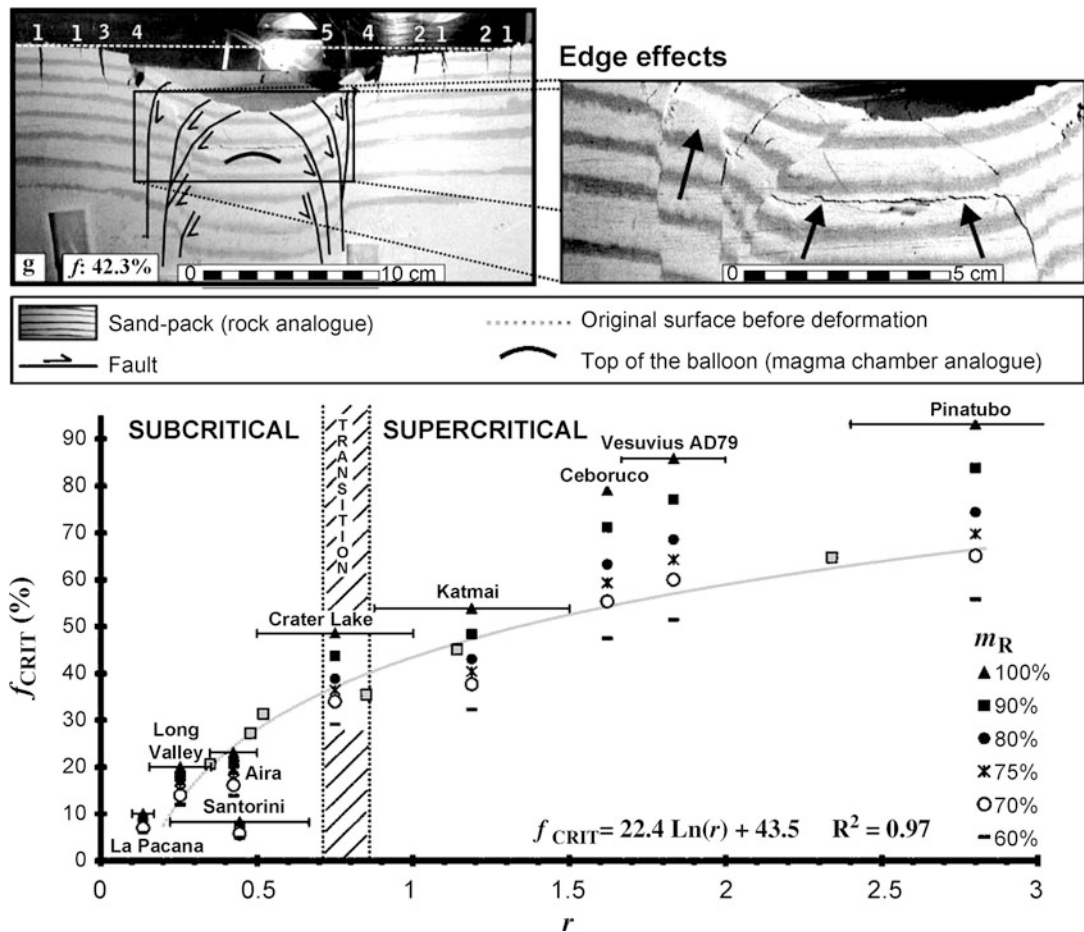


Volcanic Eruptions, Explosive: Experimental Insights, Fig. 32 Large VEI eruptions may develop into caldera collapse as the country rock overlying the magma body subsides with pressure drop in the magma. Roche and Druitt (2001) experimentally verified a model of coherent, piston-like formation of a caldera (upper left panel). Different modes of collapse were also found experimentally, some of which required significant time to elapse (upper

right panel). This suggests that the roof of a deep, narrow magma body may not collapse simultaneously with eruption, but some time afterwards. Comparison of experimentally verified theory and estimates for proportion of magma erupted (f) and roof aspect ratio (R) in natural caldera collapses (bottom panel) indicate that coherent collapse does occur. (Reprinted with kind permission from Elsevier)

fault structures and demonstrated the complex 3D nature of the faults that result from caldera collapse (Fig. 33). Fault development was monitored as the balloon was drained to establish the sequence of events prior to, and during, caldera collapse. The relationship between the proportion of water drained from the balloon to make collapse inevitable (f_{crit}) and the aspect ratio of the roof (r) is empirically given as

$f_{crit} = 22.4 \ln(r) + 43.5$ (Fig. 33). Data from caldera-forming eruptions were compared to the experimental trend and found to be in reasonable agreement. This suggests that the small-scale experiments mimic the first-order behavior of much larger-scale caldera collapse, but that there are issues of field measurement and second order process that require further investigation. Nevertheless, this work demonstrates the potential of



Volcanic Eruptions, Explosive: Experimental Insights, Fig. 33 Geyer et al. (2006) experimentally ascertained (top panel) the critical proportion of a magma chamber that must be erupted for caldera collapse to occur (f_{CRIT}). This was related to the aspect ratio of the roof (r) by an empirical equation. Experimental data (gray squares) were compared with values estimated for caldera-forming eruptions (bottom panel) for both subcritical (coherent or

piston) collapse and supercritical (non-coherent) collapse. The estimated f_{CRIT} values are a function of the percentage of magma actually removable (m_R). This work raises the possibility of being able to forecast the timing and volume of caldera collapse on the basis of previous eruption history and estimates of chamber depth and volume. (Reprinted with kind permission from Elsevier)

being able to predict the onset and volume of caldera collapse from combining field observation and laboratory experiment.

Future Directions

Experiments investigating processes that could occur both during, and as a consequence of, explosive volcanic eruptions are many and varied.

All the experiments provide insight into volcanic processes because they are all analogous in certain aspects, whether in size in the case of experiments with natural materials, or with analogue materials as well. Curiosity into the fluid dynamics of degassing and expanding multi-phase flows with non-Newtonian and viscoelastic materials will continue to expand knowledge over a wide parameter range of which volcanic processes form a small but motivating part. Experiments

will form a key part of such research and will continue to inspire numerical models that can provide a data density unavailable experimentally. One future direction could be the development of experiments that use much greater volumes of fluid and run for longer time periods. This would be especially attractive using natural materials, but the apparatus and instrumentation challenge is considerable.

The richness of phenomena that emerge from this complex system suggests that volcanoes are individuals in their eruptive behavior, an observation common to field measurement of volcanoes. Eruption models will increasingly be tested by their ability to account for observed eruption phenomena, and linking of detailed field measurement, volcano-specific experimentation and a unified modeling approach is one of the major current drivers of experimental volcanism. Highly instrumented volcanoes where measurement, in time and space, of such parameters as wide-spectrum ground and atmosphere motion, thermal outputs, chemistry and physics of magma components, electrical effects, flow rates and flow features are required to drive the understanding and forecasting of volcanic events; a task that, in general, becomes more challenging as VEI increases.

The role of experimental models is in providing a real system that can be repeated and measured in some detail. Computer modeling, verified by many experiments, may then be applied at volcanic scale. Quantitative comparison to field observation then tests the model system. Such combined tactics are likely to play a major role in the advancement of volcanology, and it is possible to envisage near real-time interpretation of field measurement in terms of fluid-dynamic, elasto-dynamic, atmospheric and chemical models built on the results of field, numerical and experimental volcanism. Such an autonomous system would be a major tool in quantifying fluid behavior in volcanic conduits and provide rapid assessment of changes taking place.

Acknowledgments We thank Bernard Chouet for a highly constructive review of this chapter, William H. K. Lee for editing this section and the staff at Springer including Julia Koerting and Kerstin Kindler.

Bibliography

Primary Literature

- Alidibirov M, Dingwell DB (1996a) Magma fragmentation by rapid decompression. *Nature* 380:146–149
- Alidibirov M, Dingwell DB (1996b) An experimental facility for the investigation of magma fragmentation by rapid decompression. *Bull Volcanol* 58: 411–416
- Alidibirov M, Dingwell DB (2000) Three fragmentation mechanisms for highly viscous magma under rapid decompression. *J Volcanol Geotherm Res* 100: 413–421
- Alidibirov M, Panov V (1998) Magma fragmentation dynamics: experiments with analogue porous low-strength material. *Bull Volcanol* 59(7):481–489
- Alidibirov M, Dingwell DB, Stevenson RJ, Hess K-U, Webb SL, Zinke J (1997) Physical properties of the 1980 Mount St. Helens cryptodome magma. *Bull Volcanol* 59:103–111
- Anilkumar AV, Sparks RSJ, Sturtevant B (1993) Geological implications and applications of high-velocity two-phase flow experiments. *J Volcanol Geotherm Res* 56(1–2):145–160
- Auger E, D’Auria L, Martini M, Chouet B, Dawson P (2006) Real-time monitoring and massive inversion of source parameters of very long period seismic signals: an application to Stromboli Volcano, Italy. *Geophys Res Lett* 33(4):L04301
- Bagdassarov NS, Dingwell DB (1992) A rheological investigation of vesicular rhyolite. *J Volcanol Geophys Res* 50:307–322
- Bagdassarov NS, Dingwell DB (1993) Frequency dependent rheology of vesicular rhyolite. *J Geophys Res* 98: 6477–6487
- Bagdassarov N, Pinkerton H (2004) Transient phenomena in vesicular lava flows based on laboratory experiments with analogue materials. *J Volcanol Geotherm Res* 132: 115–136
- Barenblatt GI (2003) *Scaling*. Cambridge University Press, Cambridge. ISBN 0-521-53394-5
- Behrens H, Gaillard F (2006) Geochemical aspects of melts: volatiles and redox behavior. *Elements* 2(5): 275–280. ISSN 1811-5209
- Berthoud G (2000) Vapor explosions. *Annu Rev Fluid Mech* 32:573–611
- Birch F, Dane EB (1942) Viscosity. In: Birch F, Schairer JF, Spicer HC (eds) *Handbook of physical constants*, vol 97. Geological Society of America Special Paper, vol 36, pp 97–173
- Blackburn EA, Wilson L, Sparks RSJ (1976) Mechanisms and dynamics of Strombolian activity. *J Geol Soc Lond* 132:429–440
- Blower JD (2001) Factors controlling permeability-porosity relationships in magma. *Bull Volcanol* 63: 497–504
- Blower JD, Keating JP, Mader HM, Phillips JC (2001) Inferring volcanic degassing processes from

- vesicle size distributions. *Geophys Res Lett* 28(2): 347–350
- Bottinga Y, Weill DF (1972) The viscosity of magmatic silicate liquids: a model for calculation. *Am J Sci* 272: 438–475
- Burgisser A, Gardner JE (2005) Experimental constraints on degassing and permeability in volcanic conduit flow. *Bull Volcanol* 67:42–56
- Burgisser A, Bergantz GW, Breidenthal RE (2005) Addressing complexity in laboratory experiments: the scaling of dilute multiphase flows in magmatic systems. *J Volcanol Geotherm Res* 141:245–265
- Burton M, Allard P, Muré F, La Spina A (2007) Magmatic gas composition reveals the source depth of slug-driven Strombolian explosive activity. *Science* 317: 227–230
- Büttner R, Roder H, Zimanowski B (1997) Electrical effects generated by experimental volcanic explosions. *Appl Phys Lett* 70(14):1903–1905
- Büttner R, Dellino P, Zimanowski B (1999) Identifying magma-water interaction from the surface features of ash particles. *Nature* 401(6754):688–690
- Büttner R, Zimanowski B, Roder H (2000) Short-time electrical effects during volcanic eruption: Experiments and field measurements. *J Geophys Res* 105(B2):2819–2827
- Cagnoli B, Barmin A, Melnik O, Sparks RSJ (2002) Depressurization of fine powders in a shock tube and dynamics of fragmented magma in volcanic conduits. *Earth Planet Sci Lett* 204(1–2):101–113
- Cambridge Polymer Group (2002) The Cambridge Polymer Group Silly Putty™ “Egg”. <http://www.campoly.com/documents/appnotes/sillyputty.pdf>. Accessed 10 Oct 2008
- Carey SN, Sigurdsson H (1982) Influence of particle aggregation on deposition of distal tephra from the May 18, 1980, eruption of Mount St-Helens volcano. *J Geophys Res* 87(NB8):7061–7072
- Carroll MR, Holloway JR (eds) (1994) Volatiles in magmas. *Rev Mineral* 30:517. ISBN 0-939950-36-7
- Chojnicki K, Clarke AB, Phillips JC (2006) A shock-tube investigation of the dynamics of gas-particle mixtures: Implications for explosive volcanic eruptions. *Geophys Res Lett* 33:L15309. <https://doi.org/10.1029/2006GL026414>
- Chouet BA, Hamisevicz B, McGetchin TR (1974) Photoballistics of volcanic jet activity at Stromboli, Italy. *J Geophys Res* 79:4961–4976
- Chouet B, Dawson P, Ohminato T, Martini M, Saccorotti G, Giudicepietro F, De Luca G, Milana G, Scarpa R (2003) Source mechanisms of explosions at Stromboli Volcano, Italy, determined from moment-tensor inversions of very-long-period data. *J Geophys Res* 108(B1):2019. <https://doi.org/10.1029/2002JB001919>
- Chouet B, Dawson P, Nakano M (2006) Dynamics of diffusive bubble growth and pressure recovery in a bubbly rhyolitic melt embedded in an elastic solid. *J Geophys Res* 111:B07310
- Chouet B, Dawson P, Martini M (2008) Shallow-conduit dynamics at Stromboli Volcano, Italy, imaged from waveform inversions. In: Lane SJ, Gilbert JS (eds) Fluid motion in volcanic conduits: a source of seismic and acoustic signals. *Geol Soc Lond Spec Publ* 307: 57–84. <https://doi.org/10.1144/SP307.5>
- Clift R, Grace JR, Weber ME (1978) Bubbles, drops and particles. Academic, London, p 380. ISBN 012176950X
- Costa A (2006) Permeability-porosity relationship: a reexamination of the Kozeny–Carman equation based on a fractal pore-space geometry assumption. *Geophys Res Lett* 33:L02318
- Dellino P, Zimanowski B, Büttner R, La Volpe L, Mele D, Sulpizio R (2007) Large-scale experiments on the mechanics of pyroclastic flows: design, engineering, and first results. *J Geophys Res* 112:B04202
- Dickinson JT, Langford SC, Jensen LC, McVay GL, Kelso JF, Pantano CG (1988) Fractoemission from fused-silica and sodium-silicate glasses. *J Vac Sci Technol A* 6(3):1084–1089
- Dingwell DB (1996) Volcanic dilemma: flow or blow? *Science* 273(5278):1054–1055
- Dingwell DB, Webb SL (1989) Structural relaxation in silicate melts and non-Newtonian melt rheology in geologic processes. *Phys Chem Miner* 16:508–516
- Donaldson EE, Dickinson JT, Bhattacharya SK (1988) Production and properties of ejecta released by fracture of materials. *J Adhes* 25(4):281–302
- Druitt TH, Avard G, Bruni G, Lettieri P, Maez F (2007) Gas retention in fine-grained pyroclastic flow materials at high temperatures. *Bull Volcanol* 69(8):881–901
- Eichelberger JC, Carrigan CR, Westrich HR, Price RH (1986) Non-explosive silicic volcanism. *Nature* 323: 598–602
- Gardner JE (2007) Heterogeneous bubble nucleation in highly viscous silicate melts during instantaneous decompression from high pressure. *Chem Geol* 236: 1–12
- Geyer A, Folch A, Marti J (2006) Relationship between caldera collapse and magma chamber withdrawal: an experimental approach. *J Volcanol Geotherm Res* 157(4):375–386
- Gilbert JS, Lane SJ (1994) The origin of accretionary lapilli. *Bull Volcanol* 56(5):398–411
- Gladstone C, Ritchie LJ, Sparks RSJ, Woods AW (2004) An experimental investigation of density-stratified inertial gravity currents. *Sedimentology* 51(4):767–789
- Gonnermann HM, Manga M (2003) Explosive volcanism may not be the inevitable consequence of magma fragmentation. *Nature* 42:432–435
- Grunewald U, Zimanowski B, Büttner R, Phillips LF, Heide K, Büchel G (2007) MFCI experiments on the influence of NaCl-saturated water on phreatomagmatic explosions. *J Volcanol Geotherm Res* 159:126–137
- Hess K-U, Dingwell DB (1996) Viscosities of hydrous leucogranitic melts: a non-Arrhenian model. *Am Mineral* 81:1297–1300

- Hill LG, Sturtevant B (1990) An experimental study of evaporation waves in a superheated liquid. In: Meier GEA, Thompson PA (eds) *Adiabatic waves in liquid-vapor systems*. IUTAM Symposium Göttingen, Germany. Springer, Berlin, pp 25–37. ISBN 3-540-50203-3
- Hoover SR, Cashman KV, Manga M (2001) The yield strength of subliquidus basalts – experimental results. *J Volcanol Geotherm Res* 107(1–3):1–18
- Hurwitz S, Navon O (1994) Bubble nucleation in rhyolitic melts: experiments at high pressure, temperature, and water content. *Earth Planet Sci Lett* 122(3–4):267–280
- Ichihara M, Rittel D, Sturtevant B (2002) Fragmentation of a porous viscoelastic material: implications to magma fragmentation. *J Geophys Res* 107(B10):2229. <https://doi.org/10.1029/2001JB000591>
- Ida Y (2007) Driving force of lateral permeable gas flow in magma and the criterion of explosive and effusive eruptions. *J Volcanol Geotherm Res* 162(3–4):172–184
- Ishibashi H, Sato H (2007) Viscosity measurements of subliquidus magmas: alkali olivine basalt from the Higashi-Matsura district, Southwest Japan. *J Volcanol Geotherm Res* 160(3–4):223–238
- James MR, Lane SJ, Gilbert JS (2000) Volcanic plume electrification: experimental investigation of a fracture-charging mechanism. *J Geophys Res* 105(B7):16641–16649
- James MR, Gilbert JS, Lane SJ (2002) Experimental investigation of volcanic particle aggregation in the absence of a liquid phase. *J Geophys Res* 107(B9):2191
- James MR, Lane SJ, Gilbert JS (2003) Density, construction, and drag coefficient of electrostatic volcanic ash aggregates. *J Geophys Res* 108(B9):2435
- James MR, Lane SJ, Chouet B, Gilbert JS (2004) Pressure changes associated with the ascent and bursting of gas slugs in liquid-filled vertical and inclined conduits. *J Volcanol Geotherm Res* 129(1–3):61–82
- James MR, Lane SJ, Chouet BA (2006) Gas slug ascent through changes in conduit diameter: laboratory insights into a volcano-seismic source process in low-viscosity magmas. *J Geophys Res* 111:B05201. <https://doi.org/10.1029/2005JB003718>
- James MR, Lane SJ, Corder SB (2008) Modelling the rapid near-surface expansion of gas slugs in low-viscosity magmas. In: Lane SJ, Gilbert JS (eds) *Fluid motion in volcanic conduits: a source of seismic and acoustic signals*. *Geol Soc Lond Spec Publ* 307:147–167. <https://doi.org/10.1144/SP307.9>
- Jaupart C, Allegre CJ (1991) Gas content, eruption rate and instabilities of eruption regime in silicic volcanoes. *Earth Planet Sci Lett* 102:413–429
- Jaupart C, Vergnolle S (1988) Laboratory models of Hawaiian and Strombolian eruptions. *Nature* 331(6151):58–60
- Jaupart C, Vergnolle S (1989) The generation and collapse of a foam layer at the roof of a basaltic magma chamber. *J Fluid Mech* 203:347–380
- Kalinichev AG (2001) Molecular simulations of liquid and supercritical water: thermodynamics, structure, and hydrogen bonding. *Rev Mineral Geochem* 42:83–129
- Kaminski E, Jaupart C (1998) The size distribution of pyroclasts and the fragmentation sequence in explosive volcanic eruptions. *J Geophys Res* 103(B12):29759–29779
- Klug C, Cashman KV (1996) Permeability development in vesiculating magmas: implications for fragmentation. *Bull Volcanol* 58:87–100
- Kouchi A, Tsuchiyama A, Sunagawa I (1986) Effect of stirring on crystallisation kinetics of basalt: texture and element partitioning. *Contrib Mineral Petrol* 93:429–438
- Kueppers U, Perugini D, Dingwell DB (2006) “Explosive energy” during volcanic eruptions from fractal analysis of pyroclasts. *Earth Planet Sci Lett* 248(3–4):800–807
- Lane SJ, Chouet BA, Phillips JC, Dawson P, Ryan GA, Hurst E (2001) Experimental observations of pressure oscillations and flow regimes in an analogue volcanic system. *J Geophys Res* 106(B4):6461–6476
- Lane SJ, Phillips JC, Ryan GA (2008) Dome-building eruptions: insights from analogue experiments. In: Lane SJ, Gilbert JS (eds) *Fluid motion in volcanic conduits: a source of seismic and acoustic signals*. Geological Society, London
- Lautze NC, Houghton BF (2007) Linking variable explosion style and magma textures during 2002 at Stromboli volcano, Italy. *Bull Volcanol* 69(4):445–460
- Liu Y, Zhang Y, Behrens H (2005) Solubility of H₂O in rhyolitic melts at low pressures and a new empirical model for mixed H₂O–CO₂ solubility in rhyolitic melts. *J Volcanol Geotherm Res* 143:219–235
- Llewellyn EW, Manga M (2005) Bubble suspension rheology and implications for conduit flow. *J Volcanol Geotherm Res* 143:205–217
- Llewellyn EW, Mader HM, Wilson SDR (2002) The rheology of a bubbly liquid. *Proc R Soc Lond A* 458:987–1016
- Lyakhovsky V, Hurwitz S, Navon O (1996) Bubble growth in rhyolitic melts: experimental and numerical investigation. *Bull Volcanol* 58:19–32
- Mader HM (1998) Conduit flow and fragmentation. *Geol Soc Lond Spec Publ* 145:51–71
- Mader HM, Zhang Y, Phillips JC, Sparks RSJ, Sturtevant B, Stolper E (1994) Experimental simulations of explosive degassing of magma. *Nature* 372(6501):85–88
- Mader HM, Phillips JC, Sparks RSJ, Sturtevant B (1996) Dynamics of explosive degassing of magma: Observations of fragmenting two-phase flows. *J Geophys Res Solid Earth* 101(B3):5547–5560
- Mader HM, Brodsky EE, Howard D, Sturtevant B (1997) Laboratory simulations of sustained volcanic eruptions. *Nature* 388(6641):462–464
- Marshall JR, Sauke TB, Cuzzi JN (2005) Microgravity studies of aggregation in particulate clouds. *Geophys Res Lett* 32(11):L11202

- Martel C, Dingwell DB, Spieler O, Pichavant M, Wilke M (2000) Fragmentation of foamed silicic melts: an experimental study. *Earth Planet Sci Lett* 178:47–58
- Martel C, Dingwell DB, Spieler O, Pichavant M, Wilke M (2001) Experimental fragmentation of crystal- and vesicle-bearing silicic melts. *Bull Volcanol* 63:398–405
- Mason RM, Starostin AB, Melnik OE, Sparks RSJ (2006) From Vulcanian explosions to sustained explosive eruptions: the role of diffusive mass transfer in conduit flow dynamics. *J Volcanol Geotherm Res* 153(1–2):148–165
- McGetchin TR, Settle M, Chouet B (1974) Cinder cone growth modeled after Northeast Crater, Mount Etna, Sicily. *J Geophys Res* 79:3257–3272
- McMillan PF (1994) Water solubility and speciation models. In: Carroll MR, Holloway JR (eds) *Volatiles in magmas*
- Melnik O, Sparks RSJ (2002) Dynamics of magma ascent and lava extrusion at Soufrière Hills Volcano, Montserrat. In: Druitt T, Kokelaar P (eds) *The eruption of Soufrière Hills Volcano, Montserrat, from 1995 to 1999*. Geological Society, London, pp 153–171
- Moore JG, Peck DL (1962) Accretionary lapilli in volcanic rocks of the western continental United-States. *J Geol* 70(2):182–193
- Mourtada-Bonnefoi CC, Mader HM (2001) On the development of highly-viscous skins of liquid around bubbles during magmatic degassing. *Geophys Res Lett* 28(8):1647–1650
- Mourtada-Bonnefoi CC, Mader HM (2004) Experimental observations of the effects of crystals and pre-existing bubbles on the dynamics and fragmentation of vesiculating flows. *J Volcanol Geotherm Res* 129:83–97. [https://doi.org/10.1016/S0377-0273\(03\)00233-6](https://doi.org/10.1016/S0377-0273(03)00233-6)
- Murase T (1962) Viscosity and related properties of volcanic rocks at 800° to 1400°C. *J Fac Sci Hokkaido Univ Ser VII* 1:487–584
- Murase T, McBirney AR (1973) Properties of some common igneous rocks and their melts at high-temperatures. *Geol Soc Am Bull* 84(11):3563–3592
- Namiki A, Manga M (2005) Response of a bubble bearing viscoelastic fluid to rapid decompression: implications for explosive volcanic eruptions. *Earth Planet Sci Lett* 236:269–284
- Namiki A, Manga M (2006) Influence of decompression rate on the expansion velocity and expansion style of bubbly fluids. *J Geophys Res* 111:B11208
- Newhall CG, Self S (1982) The explosivity index (VEI) – an estimate of explosive magnitude for historical volcanism. *J Geophys Res* 87(NC2):1231–1238
- Parfitt EA (2004) A discussion of the mechanisms of explosive basaltic eruptions. *J Volcanol Geotherm Res* 134:77–107
- Phillips JC, Lane SJ, Lejeune A-M, Hilton M (1995) Gum rosin – acetone system as an analogue to the degassing behaviour of hydrated magmas. *Bull Volcanol* 57: 263–268
- Pinkerton H, Norton G (1995) Rheological properties of basaltic lavas at sub-liquidus temperatures – laboratory and field-measurements on lavas from Mount Etna. *J Volcanol Geotherm Res* 68(4):307–323
- Pyle DM (2000) Sizes of volcanic eruptions. In: Sigurdsson H, Houghton B, McNutt SR, Rymer H, Stix J (eds) *Encyclopaedia of volcanoes*. Academic, pp 263–269. ISBN 0-12-643140-X
- Riedel C, Ernst GGJ, Riley M (2003) Controls on the growth and geometry of pyroclastic constructs. *J Volcanol Geotherm Res* 127(1–2):121–152
- Riley CM, Rose WI, Bluth GJS (2003) Quantitative shape measurements of distal volcanic ash. *J Geophys Res Solid Earth* 108(B10):2504
- Ripepe M, Ciliberto S, Della Schiava M (2001) Time constraints for modeling source dynamics of volcanic explosions at Stromboli. *J Geophys Res Solid Earth* 106(B5):8713–8727
- Roche O, Druitt TH (2001) Onset of caldera collapse during ignimbrite eruptions. *Earth Planet Sci Lett* 191(3–4):191–202
- Saar MO, Manga M (1999) Permeability-porosity relationship in vesicular basalts. *Geophys Res Lett* 26: 111–114
- Schumacher R, Schmincke HU (1995) Models for the origin of accretionary lapilli. *Bull Volcanol* 56(8): 626–639
- Seyfried R, Freundt A (2000) Experiments on conduit flow and eruption behavior of basaltic volcanic eruptions. *J Geophys Res Solid Earth* 105:23727–23740
- Shaw HR (1972) Viscosities of magmatic silicate liquids: an empirical method of prediction. *Am J Sci* 272: 870–893
- Sparks RSJ (1997) Causes and consequences of pressurisation in lava dome eruptions. *Earth Planet Sci Lett* 150:177–189
- Sparks RSJ, Bursik MI, Carey SN, Gilbert JS, Glaze LS, Sigurdsson H, Woods AW (1997) *Volcanic plumes*. Wiley, Chichester. ISBN 0-471-93901-3
- Spera FJ, Borgia A, Strimple J, Feigenson M (1988) Rheology of melts and magmatic suspensions I. Design and calibration of a concentric cylinder viscometer with application to rhyolitic magma. *J Geophys Res* 93: 10273–10294
- Spieler O, Alidibirov M, Dingwell DB (2003) Grain-size characteristics of experimental pyroclasts of 1980 Mount St. Helens cryptodome dacite: effects of pressure drop and temperature. *Bull Volcanol* 65:90–104
- Spieler O, Kennedy B, Kueppers U, Dingwell DB, Scheu B, Taddeucci J (2004) The fragmentation threshold of pyroclastic rocks. *Earth Planet Sci Lett* 226: 139–148
- Sugioka I, Bursik M (1995) Explosive fragmentation of erupting magma. *Nature* 373(6516):689–692
- Suzuki T (1983) A theoretical model for dispersion of tephra. In: Shimozuru D, Yokoyama I (eds) *Arc volcanism, physics and tectonics*. Terra Scientific Publishing Company (Terrapub), Tokyo, pp 95–113
- Taddeucci J, Spieler O, Kennedy B, Pompilio M, Dingwell DB, Scarlato P (2004) Experimental and analytical modeling of basaltic ash explosions at Mount Etna, Italy, 2001. *J Geophys Res* 109:B08203. <https://doi.org/10.1029/2003JB002952>
- Taddeucci J, Spieler O, Ichihara M, Dingwell DB, Scarlato P (2006) Flow and fracturing of viscoelastic media

- under diffusion-driven bubble growth: an analogue experiment for eruptive volcanic conduits. *Earth Planet Sci Lett* 243:771–785
- Taddeucci J, Scarlato P, Andronico D, Cristaldi A, Büttner R, Zimanowski B, Küppers U (2007) Advances in the study of volcanic ash. *Eos Trans AGU* 88(24):253
- Takeuchi S, Nakashima S, Tomiya A, Shinohara H (2005) Experimental constraints on the low gas permeability of vesicular magma during decompression. *Geophys Res Lett* 32:L10312
- Takeuchi S, Nakashima S, Tomiya A (2008) Permeability measurements of natural and experimental volcanic materials with a simple permeameter: Toward an understanding of magmatic degassing process. *J Volcanol Geotherm Res*. <https://doi.org/10.1016/j.jvolgeores.2008.05.010>
- Textor C, Graf HF, Longo A, Neri A, Ongaro TE, Papale P, Timmreck C, Ernst GG (2005) Numerical simulation of explosive volcanic eruptions from the conduit flow to global atmospheric scales. *Ann Geophys* 48(4–5):817–842
- Trigila R, Battaglia M, Manga M (2007) An experimental facility for investigating hydromagmatic eruptions at high-pressure and high-temperature with application to the importance of magma porosity for magma-water interaction. *Bull Volcanol* 69:365–372
- Tuffen H, Dingwell D (2005) Fault textures in volcanic conduits: evidence for seismic trigger mechanisms during silicic eruptions. *Bull Volcanol* 67:370–387
- Walker GPL, Wilson L, Howell ELG (1971) Explosive volcanic eruptions. 1. Rate of fall of pyroclasts. *Geophys J Roy Astron Soc* 22(4):377–383
- Watson EB (1994) Diffusion in volatile-bearing magmas. In: Carroll MR, Holloway JR (eds) *Volatiles in magmas*
- Wilson L, Huang TC (1979) Influence of shape on the atmospheric settling velocity of volcanic ash particles. *Earth Planet Sci Lett* 44(2):311–324
- Wohletz KH (1983) Mechanisms of hydrovolcanic pyroclast formation: grain-size, scanning electron microscopy, and experimental studies. *J Volcanol Geotherm Res* 17(1–4):31–63
- Woolley AR, Church AA (2005) Extrusive carbonatites: a brief review. *Lithos* 85(1–4):1–14
- Yokoyama I (2005) Growth rates of lava domes with respect to viscosity of magmas. *Ann Geophys* 48(6):957–971
- Zhang YX (1998) Experimental simulations of gas-driven eruptions: kinetics of bubble growth and effect of geometry. *Bull Volcanol* 59(4):281–290
- Zhang Y (1999) A criterion for the fragmentation of bubbly magma based on brittle failure theory. *Nature* 402:648–650
- Zhang Y, Behrens H (2000) H₂O diffusion in rhyolitic melts and glasses. *Chem Geol* 169:243–226
- Zhang YX, Sturtevant B, Stolper EM (1997) Dynamics of gas-driven eruptions: experimental simulations using CO₂-H₂O-polymer system. *J Geophys Res Solid Earth* 102(B2):3077–3096
- Zhang Y, Xu Z, Liu Y (2003) Viscosity of hydrous rhyolitic melts inferred from kinetic experiments: a new viscosity model. *Am Mineral* 88:1741–1752
- Zimanowski B, Lorenz V, Frohlich G (1986) Experiments on phreatomagmatic explosions with silicate and carbonatitic melts. *J Volcanol Geotherm Res* 30(1–2):149–153
- Zimanowski B, Büttner R, Lorenz V, Häfele H-G (1997) Fragmentation of basaltic melt in the course of explosive volcanism. *J Geophys Res* 102(B1):803–814

Books and Reviews

- Barnes HA (1997) Thixotropy – a review. *J Non-Newtonian Fluid Mech* 70:1–33
- Barnes HA (1999) The yield strength – a review or ‘πανταρὰ’ everything flows? *J Non-Newtonian Fluid Mech* 81:133–178
- Chhabra RP (2006) Bubbles, drops, and particles in non-Newtonian fluids. CRC Press, London, p 771. ISBN 0824723295
- Chouet B (2003) Volcano seismology. *Pure Appl Geophys* 160:739–788
- De Kee D, Chhabra RP (2002) Transport processes in bubbles, drops, and particles. Taylor and Francis, p 352. ISBN 1560329068
- Fan L-S, Zhu C (1998) Principles of gas-solid flows. Cambridge University Press, Cambridge, p 575. ISBN 0521581486
- Gilbert JS, Sparks RSJ (1998) The physics of explosive volcanic eruptions. The Geological Society, London. ISBN 1-86239-020-7
- Gonnermann HM, Manga M (2007) The fluid mechanics inside a volcano. *Annu Rev Fluid Mech* 39:321–356
- Henderson G, Calas G, Stebbins J (eds) (2006) Glasses and melts: linking geochemistry and materials science. *Elements* 2(5):257–320. ISSN 1811–5209
- Kaminski E, Tait S, Carazzo G (2005) Turbulent entrainment in jets with arbitrary buoyancy. *J Fluid Mech* 526:361–376
- Mather TA, Harrison RG (2006) Electrification of volcanic plumes surveys. *Geophysics* 71(4):387–432
- Moretti R, Richet P, Stebbins JF (2006) Physics, chemistry and rheology of silicate melts and glasses. *Chem Geol* 229(1–3):1–226
- Parfitt L, Wilson L (2008) Fundamentals of physical volcanology. Blackwell, p 256. ISBN 0632054433
- Sigurdsson H, Houghton B, McNutt SR, Rymer H, Stix J (eds) (2000) *Encyclopaedia of volcanoes*. Academic. ISBN 0-12-643140-X
- Tong L-S, Tang YS (1997) Boiling heat transfer and two-phase flow. Taylor and Francis, p 542. ISBN 1560324856
- Webb S (1997) Silicate melts: relaxation, rheology, and the glass transition. *Rev Geophys* 35(2):191–218
- Wood CA (1980) Morphometric evolution of cinder cones. *J Volcanol Geotherm Res* 7:387–413

Volcanic Eruptions: Cyclicity During Lava Dome Growth

Oleg Melnik¹, R. Stephen J. Sparks²,
Antonio Costa² and Alexei A. Barmin¹

¹Institute of Mechanics, Moscow State
University, Moscow, Russia

²Earth Science Department, University of Bristol,
Bristol, UK

Article Outline

Glossary

Definition of the Subject

Introduction

Dynamics of Magma Ascent During Extrusive
Eruptions

Short-Term Cycles

Long-Term Cycles

Future Directions

Bibliography

Glossary

Andesite Magma or volcanic rock that is characterized by intermediate SiO₂ concentration. Andesite magmas have rheological properties that are intermediate between basalt and rhyolite magmas. Silica content in andesites ranges from approximately 52 to 66 weight percent. Common minerals in andesite include plagioclase, amphibole, and pyroxene. Andesite is typically erupted at temperatures between 800 °C and 1000 °C. Andesite is particularly common in subduction zones, where tectonic plates converge and water is introduced into the mantle.

Basalt Magma or volcanic rock that contains not more than about 52% SiO₂ by weight. Basaltic magmas have a low viscosity. Volcanic gases can escape easily without generating high

eruption columns. Basalt is typically erupted at temperatures between 1100 °C and 1250 °C. Basalt flows cover about 70% of the Earth's surface and huge areas of the terrestrial planets and so are the most important of all crustal igneous rocks.

Bingham liquid A fluid that does not flow in response to an applied stress until a critical yield stress is reached. Above the critical yield stress, strain rate is proportional to the applied stress, as in a Newtonian fluid.

Bubbly flow A multi-phase flow regime, in which the gas phase appears as bubbles suspended in a continuous liquid phase.

Conduit A channel, through which magma flows toward the Earth's surface. Volcanic conduits can commonly be approximately cylindrical and typically a few tens of meters across or bounded by near-parallel sides in a magma-filled fracture. Conduits can be vertical or inclined.

Crystallization Conversion, partial or total, of a silicate melt into crystals during solidification of magma.

Degassing n. (degas v.) The process by which volatiles that are dissolved in silicate melts come out of solution in the form of bubbles. Open- and closed-system degassing can be distinguished. In the former, volatiles can be lost or gained by the system. In the latter, the total amount of volatiles in the bubbles and in solution in the magma is conserved.

Differentiation The process of changing the chemical composition of magma by processes of crystallization accompanied by separation of melts from crystals.

Dome A steep-sided, commonly bulbous, extrusion of lava or shallow intrusion (cryptodome). Domes are commonly, but not exclusively, composed of SiO₂-rich magmas. In dome-forming eruptions, the erupted magma is so viscous, or the discharge rate so slow, that lava accumulates very close to the vent region, rather than flowing away. Pyroclastic flows can

be generated by collapse of lava domes. Recent eruptions producing lava domes include the 1995–2006 eruption of the Soufrière Hills Volcano, Montserrat, and the 2004–2006 eruption of Mount St. Helens, USA.

Dyke A sheet-like igneous intrusion, commonly vertical or near vertical, which cuts across pre-existing, older, geological structures. During magmatism, dykes transport magma toward the surface or laterally in fracture-like conduits. In the geologic record, dykes are preserved as sheetlike bodies of igneous rocks.

Explosive eruption A volcanic eruption in which gas expansion tears the magma into numerous fragments with a wide range of sizes. The mixture of gas and entrained fragments flows upward and outward from volcanic vents at high speed into the atmosphere. Depending on the volume of erupted material, eruption intensity, and sustainability, explosive eruptions are classified as Strombolian, Vulcanian, sub-Plinian, Plinian, or Mega-Plinian; this order is approximately in the order of increasing intensity. Strombolian and Vulcanian eruptions involve very short-lived explosions.

Extrusive flow or eruption A nonexplosive (non-pyroclastic) magma flow from a volcanic conduit during a lava dome-building eruption or lava flow.

Mafic Magma, lava, or tephra with silica concentrations of approximately $\text{SiO}_2 < 55\%$.

Magma chamber A subsurface volume within which magma accumulates, differentiates, and crystallizes. Igneous intrusions can constrain the form and size of some magma chambers, but in general the shape and volume of magma chambers beneath active volcanoes are poorly known. Magma reservoir is an equivalent term.

Magma Molten rock that consists of up to three components: liquid silicate melt, suspended crystalline solids, and gas bubbles. It is the raw material of all volcanic processes. Silicate magmas are the most common magma type and consist of long, polymeric chains and rings of Si-O tetrahedra, between which are located cations (e.g., Ca^{2+} , Mg^{2+} , Fe^{2+} , and Na^+). Anions (e.g., OH^- , F^- , Cl^- , and S^-)

can substitute for the oxygen in the silicate framework. The greater the silica (SiO_2) content of the magma, the more chains and rings of silicate tetrahedra there are to impede each other, and hence the viscosity of the magma increases. The pressure regime and composition of the magma control the minerals that nucleate and crystallize from a magma when it cools or degasses.

Melt Liquid part of magma. Melts (usually silicate) contain variable amounts of dissolved volatiles. The primary volatiles are usually water and carbon dioxide.

Microlite Crystal with dimensions less than 100 μm . Usually microlites crystallize at shallow levels of magmatic system.

Newtonian liquid A liquid for which the strain rate is proportional to the applied stress. The proportionality coefficient is called the viscosity.

Phenocryst Crystal with dimensions larger than 100 μm . Usually phenocrysts grow in magmatic reservoirs prior to an eruption and/or are entrained by magma in the chamber.

Pyroclastic flow or surge A gas-particle flow of pyroclasts suspended in a mixture of hot air, magmatic gas, and fine ash. The flow originates by the gravitational collapse of a dense, turbulent explosive eruption column at the source vent or by dome collapse and moves downslope as a coherent flow. Pyroclastic flows and surges are distinguished by particle concentration in the flow, surges being more dilute. Variations in particle concentration result in differences in the deposits left by flows and surges.

Silicic Magma, lava, or tephra with silica concentrations of approximately $\text{SiO}_2 > 55\%$. The magmas are commonly rich in Al-, Na-, and K-bearing minerals. Silicic magmas are typically very viscous and can have high volatile contents. Rhyolite is an example of a silicic magma.

Volatile A component in a magmatic melt which can be partitioned in the gas phase in significant amounts during some stage of magma history. The most common volatile in magmas is water vapor H_2O , but there are commonly

also significant quantities of CO₂, SO₂, and halogens.

Definition of the Subject

We consider the process of slow extrusion of very viscous magma that forms lava domes. Dome-building eruptions are commonly associated with hazardous phenomena, including pyroclastic flows generated by dome collapses, explosive eruptions, and volcanic blasts. These eruptions commonly display fairly regular alternations between periods of high and low or no activity with timescales from hours to years. Usually hazardous phenomena are associated with periods of high magma discharge rate; thus, understanding the causes of pulsatory activity during extrusive eruptions is an important step toward forecasting volcanic behavior, especially the transition to explosive activity when magma discharge rate increases by a few orders of magnitude. In recent years the risks have increased because the population density in the vicinity of many active volcanoes has increased.

Introduction

Many volcanic eruptions involve the formation of lava domes, which are extrusions of very viscous, degassed magmas. The magma is so viscous that it accumulates close to the vent. Extrusion of lava domes is a slow and long-lived process and can continue for many years or even decades (Newhall and Melson 1983; Sparks 1997; Sparks and Aspinall 2004). Typical horizontal dimensions of lava domes are several hundred meters, heights are of an order of tens to several hundred meters, and volumes are several million to hundreds of million cubic meters. Typical magma discharge rates (measured as the increase of dome volume with time in dense rock equivalent (DRE)) can reach up to 20–40 m³/s but are usually below 10 m³/s (Sparks 1997).

Dome-building eruptions are commonly associated with hazardous phenomena, including pyroclastic flows and tsunamis generated by

dome collapses, explosive eruptions, and volcanic blasts. Dome-building eruptions can also contribute to edifice instability and sector collapse, as occurred on Montserrat on 26 December 1997 (Sparks and Young 2002). Lava dome activity can sometimes precede or follow major explosive eruptions; the eruption of Pinatubo, Philippines (1991), is an example of the former (Hoblitt et al. 1996), and the eruption of Mount St. Helens, USA (1980–1986), is an example of the latter (Swanson and Holcomb 1990).

Several lava dome eruptions have been documented in detail and show quite complex behaviors. Substantial fluctuations in magma discharge rate have been documented. In some cases these fluctuations can be quite regular (nearly periodic), as in the extrusion of lava in 1980–1982 on Mount St. Helens (Swanson and Holcomb 1990) and in the 1922–2002 activity of the Santiaguito lava dome, Guatemala (Harris et al. 2002). In these cases, periods of high magma discharge rate alternate with longer periods of low magma discharge rate or no extrusion. In some volcanoes, such as Shiveluch, Kamchatka, the intervals of no extrusion are so long compared with the periods of dome growth that the episodes of dome growth have been described as separate eruptions of the volcano rather than episodes of the same eruption. Other dome-building activity can be nearly continuous and relatively steady, as observed at Mount St. Helens in 1983 (Swanson and Holcomb 1990) and at the Soufrière Hills Volcano, Montserrat, between November 1999 and July 2003. In yet other cases, the behavior can be more complex with quite sudden changes in magma discharge rate, which cannot be related to any well-defined regularity or pattern (e.g., Lascar volcano, Chile (Matthews et al. 1997)).

Pauses during lava dome-building eruptions are quite common. For example, at Mount St. Helens, there were 9 pulses of dome growth with a period of ~74 days, a duration of 1–7 days, and no growth in between (Swanson and Holcomb 1990). The Soufrière Hills Volcano, Montserrat, experienced a long (20 months) pause in extrusion after the first episode of growth (Norton et al. 2002). On Shiveluch volcano in

Kamchatka, episodes of dome growth occurred in 1980, 1993, and 2000, following a major explosion in 1964 (Fedotov et al. 2001). Each episode of dome growth began with magma discharge rate increasing over the first few weeks to a peak of 8–15 m³/s, with a gradual decline in magma discharge rate over the following year. In between the episodes very minimal activity was recorded.

Fluctuations in magma discharge rate have been documented on a variety of timescales from both qualitative and quantitative observations. Several lava dome eruptions are characterized by extrusion of multiple lobes and flow units (Nakada et al. 1999; Watts et al. 2002). In the case of the Soufrière Hills Volcano, extrusion of shear lobes can be related to spurts in discharge rate and is associated with other geophysical changes, such as onset of seismic swarms and marked changes in temporal patterns of ground tilt (Voight et al. 1998, 1999; Watts et al. 2002). These spurts in discharge rate have been fairly regular for substantial periods, occurring every 6 to 7 weeks over a 7-month period in 1997 (Costa et al. 2007b; Sparks and Young 2002; Voight et al. 1999). These spurts are commonly associated with large dome collapses and pyroclastic flows and, in some cases, with the onset of periods of repetitive Vulcanian explosions (Cole et al. 2002; Druitt et al. 2002). Consequently the recognition of this pattern has become significant for forecasting activity for hazard-assessment purposes. At Soufrière Hills Volcano and Mount Pinatubo, much shorter fluctuations in magma discharge rate have been recognized from cyclic variations in seismicity, ground tilt, gas fluxes, and rockfall activity (Denlinger and Hoblitt 1999; Voight et al. 1999; Watson et al. 2000). This cyclic activity has typical periods in the range of 4–36 h. Cyclic activity has been attributed to cycles of gas pressurization and depressurization with surges in dome growth related to degassing, rheological stiffening, and stick-slip behavior (Denlinger and Hoblitt 1999; Lensky et al. 2007; Melnik and Sparks 1999; Voight et al. 1999; Wylie et al. 1999).

Dome eruptions can show transitions to explosive activity, which sometimes can be linked to spurts in magma discharge rate. For example, in

1980, periodic episodes of lava dome extrusion on Mount St. Helens were initiated by explosive eruptions, which partly destroyed the dome that had been extruded in each previous extrusion episode (Swanson and Holcomb 1990). At Unzen volcano, Japan, a single Vulcanian explosive eruption occurred in June 1991 when the magma discharge rate was at its highest (Nakada et al. 1999). At the Soufrière Hills Volcano, repetitive series of Vulcanian explosions have occurred following large dome collapses in periods when magma discharge rates were the highest of the eruption (Druitt et al. 2002; Sparks et al. 1998). In the case of Lascar volcano, Chile, an intense Plinian explosive eruption occurred on 18 and 19 April 1993, after 9 years of dome extrusion and occasional short-lived Vulcanian explosions (Matthews et al. 1997).

Lava dome eruptions require magma with special physical properties. In order to produce a lava dome rather than a lava flow, the viscosity of the magma must be extremely high so that the lava cannot flow easily from the vent. High viscosity is a consequence of factors such as relatively low temperature (typically 750–900 °C), melt compositions rich in network-forming components (principally Si and Al), efficient gas loss during magma decompression, and crystallization as a response to cooling and degassing. Viscosities of silica-rich magmas, such as rhyolites and some andesites, are increased by several orders of magnitude by the loss of dissolved water during decompression. Many, but not all, domes also have high crystal content (up to 60 to 95 vol%), with crystallization being triggered mostly by degassing (Cashman and Blundy 2000; Sparks et al. 2000). In order to avoid fragmentation that leads to an explosive eruption, magma must have lost gas during ascent. Consider, for example, a magma at 150 MPa containing 5 wt% of dissolved water decompressed to atmospheric pressure. Without gas loss the volume fraction of bubbles will be more than 99%. Typical dome rock contains less than 20 vol% of bubbles, although there is evidence that magma at depth can be more bubble-rich (e.g., Clarke et al. 2007; Robertson et al. 1998). On the other hand, very commonly there is no change in temperature or bulk magma

composition in the products of explosive and extrusive eruptions for a particular volcano. This suggests that the properties of magma that are conducive to the formation of lava domes are controlled by physicochemical transformations that occurred during magma ascent to the surface.

Two other important factors that influence whether lava domes or flows form are topography and discharge rate. The same magma can form a dome if the discharge rate is low and a lava flow if the rate is high (Fink and Griffiths 1990; Walker 1973). The discharge rate is controlled by overall conduit resistance that is a function of viscosity, conduit size and shape, and driving pressure (the difference between chamber pressure and atmospheric pressure). Additionally the same magma can form a dome on low slopes, such as a flat crater (e.g., the mafic andesite dome of the Soufrière Volcano, St. Vincent; Huppert et al. 1982) and a lava flow on steep slopes.

Prior to an eruption, magma is usually stored in a shallow crustal reservoir called a magma chamber. For several volcanoes magma chambers can be detected and characterized by earthquake locations, seismic tomography, petrology, or interpretation of ground deformation data (Marsh 2000). Typical depths of magma chambers range from a few kilometers to tens of kilometers. Volumes range from less than 1 to several thousand km³ (Marsh 2000) but are usually less than a hundred km³. Magma chambers are connected to the surface by magma pathways called conduits. There is evidence that the conduits that feed lava dome eruptions can be both dykes and cylindrical. Dykes of a few meters width are commonly observed in the interior of eroded andesite volcanoes. Dyke feeders to lava domes have been intersected by drilling at Inyo Crater, California, USA (Mastin and Pollard 1988), and at Mount Unzen (Nakada and Eichelberger 2004). Geophysical studies point to dyke feeders; for example, fault plane solutions of shallow volcano tectonic earthquakes indicate pressure fluctuations in dykes (Roman 2005; Roman et al. 2006). Deformation data at Unzen, combined with structural analysis, indicate that the 1991–1995 dome was fed by a dyke (Nakada et al. 1999). Dykes are also the only viable

mechanism of developing a pathway through brittle crust from a deep magma chamber to the surface in the initial stages of an eruption (Lister and Kerr 1991; Rubin 1995).

Cylindrical conduits commonly develop during lava dome eruptions. The early stages of lava dome eruptions frequently involve phreatic and phreatomagmatic explosions that create near-surface craters and cylindrical conduits (Christiansen and Peterson 1981; Ohba and Kitade 2005; Sparks and Young 2002; Swanson and Holcomb 1990; Williams and Self 1983; Yokoyama et al. 1981). These explosions are usually attributed to interaction of magma rising along a dyke with groundwater. Cylindrical conduits formed by explosions are confined to relatively shallow parts of the crust, probably of order hundreds of meters in depth and <1 km, as indicated by mineralogical studies (Ohba and Kitade 2005). Examples of such initial conduit-forming activity include Mount Usu (Japan), Mt. St. Helens, and Soufrière Hills Volcano (Christiansen and Peterson 1981; Sparks and Young 2002; Yokoyama et al. 1981). Many lava dome eruptions are also characterized by Vulcanian, sub-Plinian, and even Plinian explosive eruptions. Examples include Mount Unzen, Mount St. Helens, Santiaguito, and Soufrière Hills Volcano (Nakada et al. 1999; Sparks and Young 2002; Swanson and Holcomb 1990; Williams and Self 1983). Here the fragmentation front may reach to depths of several kilometers (Mason et al. 2006) with the possibility of cylindrical conduit development due to severe underpressurization and mechanical disruption of conduit wall rocks. Subsequently domes can be preferentially fed along the cylindrical conduits created by earlier explosive activity. On the Soufrière Hills Volcano, early dome growth was characterized by extrusion of spines with nearly cylindrical shape (Sparks and Young 2002).

Observations of magma discharge rate variations on a variety of timescales highlight the need to understand the underlying dynamic controls. Research has increasingly focused on modeling studies of conduit flow dynamics during lava dome eruptions. We will restrict our discussions here to mechanisms that lead to cyclic and

quasiperiodic fluctuations in magma discharge rate on various timescales, mainly focusing on long-term cycles. Issues concerning the transition between explosive and extrusive activity are discussed in detail in (Jaupart and Allegre 1991; Slezin 1984, 2003; Woods and Koyaguchi 1994) and in the special volume of *Journal of Volcanology and Geothermal Research* dedicated to modeling of explosive eruptions (Sahagian 2005). Several papers consider the processes that occur at the surface and relate dome morphology and dimensions with controlling parameters.

Combined theoretical, experimental, and geological studies identify four main types of dome: spiny, lobate, platy, and axisymmetric (Blake 1990; Fink and Griffiths 1990, 1998). These types of dome reflect different regimes which are controlled by discharge rates, cooling rates and yield strength, and the viscosity of the dome-forming material. In recent years, mathematical modeling has been used to semiquantitatively describe spreading of lava domes, including models based on the thin layer approximation (Balmforth et al. 2001, 2006) and fully 2D simulations of lava dome growth which account for viscoelastic and viscoplastic rheologies (Hale et al. 2007; Hale and Wadge 2003).

Dynamics of Magma Ascent During Extrusive Eruptions

In order to understand the causes of cyclic behavior during extrusive eruptions, first we need to consider the underlying dynamics of volcanic systems and discuss physical and chemical transitions during magma ascent.

The physical framework for the model of a volcanic system is shown in Fig. 1. Magma is stored in a chamber at depth L , with a chamber pressure P_{ch} that is higher than hydrostatic pressure of magma column and drives magma ascent. Magma contains silicate melt, crystals, and dissolved and possibly exsolved volatiles. During ascent of the magma up the conduit, the pressure decreases, and volatiles exsolve forming bubbles. As the bubble concentration becomes substantial, bubble coalescence takes place, and permeability

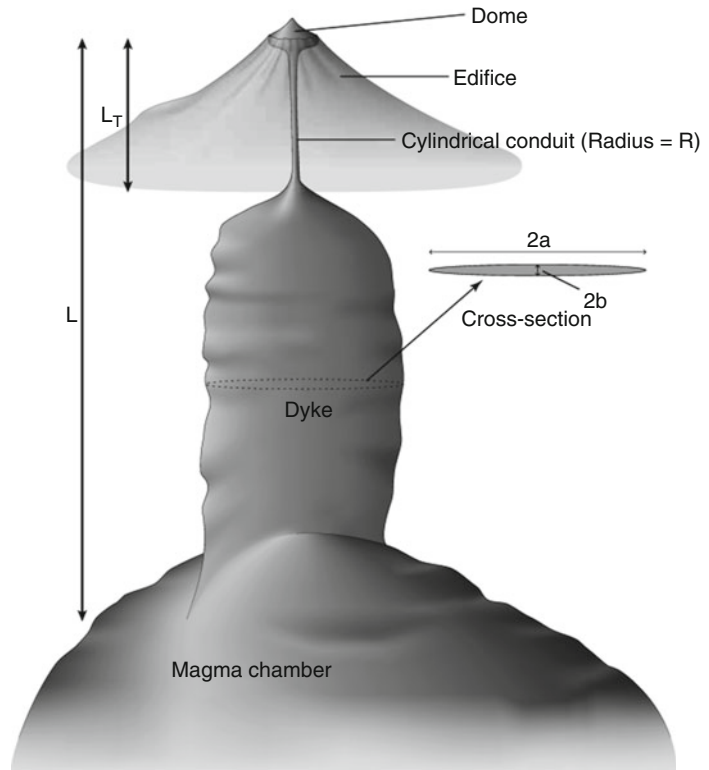
develops (Eichelberger et al. 1986; Jaupart and Allegre 1991), allowing gas to escape from ascending magma, both in vertical (through the magma) and horizontal (to conduit wall rocks) directions. If magma ascends slowly, gas escape results in a significant reduction of the volume fraction of bubbles; such a process is termed open-system degassing. In comparison closed-system degassing is characterized by a negligible gas escape. Reduction in bubble content, combined with relatively low gas pressures and efficient decoupling of the gas and melt phases, prevents magma fragmentation and, thus, development of explosive eruption (Melnik 2000; Melnik et al. 2005; Sparks 1978).

Due to typical low ascent velocities (from millimeters to a few centimeters per second), magma ascent times to the Earth's surface from the magma chamber range from a several hours to many weeks. These ascent times are often comparable with the times that are required for crystals to grow significantly and for heat exchange between magma and wall rocks. The main driving force for crystallization is related to exsolution of volatiles from the magma, leading to increase in the liquidus temperature T_L , and development of magma undercooling $\Delta T = T_L - T$ (Cashman and Blundy 2000). Crystallization leads to release of latent heat, and magma temperature can increase with respect to the initial temperature (Blundy et al. 2006). As a consequence of increasing crystal content, magma viscosity increases by several orders of magnitude (Costa 2005; Costa et al. 2007a,b), and magma becomes a non-Newtonian fluid (Saar et al. 2001). As will be shown later, crystallization induced by degassing can become a key process in causing variable flow rates.

Due to the long duration of extrusive eruptions, the magma chamber can be replenished with significant amounts of new magma from underlying sources (Huppert and Woods 2002; Murphy et al. 2000). Replenishment can lead to pressure buildup in the magma chamber and volatile and heat exchange between host and new magmas. The composition of the magma can also change over time, due to differentiation, crustal rock assimilation, or magma mixing. Thus, any model that explains magma ascent dynamics needs to

Volcanic Eruptions: Cyclicity During Lava Dome Growth,

Fig. 1 Schematic view of the volcanic system. In the upper part the conduit is cylindrical with a radius R . A transition from the cylinder to a dyke occurs at depth L_T . The length scale for the transition from cylinder to dyke is w_T . The dyke has an elliptical cross section with semiaxis lengths a_0 and b_0 . The chamber is located at depth L . In the text we used also the following auxiliary variables: $D = 2R$ for conduit diameter, $L_d = L - L_T$ for the dyke vertical length, $W_d = 2a_0$ for the dyke width, and $H_d = 2b_0$ for the dyke thickness. (After Costa et al. 2007a)



deal with many complexities. Of course there is no single model that can take into account all physical processes in a volcanic system. Additional complications arise from the fact that the physical properties of magma at high crystal contents, such as rheology or crystal growth kinetics and geometry of volcanic systems, are typically poorly constrained. Several issues regarding the dynamics of multiphase systems have not been resolved theoretically, especially for cases where the volume fractions of components of the multiphase system are comparable.

Below we will present a review of existing models that treat cyclic behavior during extrusive eruptions on different timescales.

Short-Term Cycles

Cyclic patterns of seismicity, ground deformation, and volcanic activity (Fig. 2 from Denlinger and Hoblitt 1999) have been documented at Mount Pinatubo, Philippines, in 1991 (Hoblitt et al.

1996) and Soufrière Hills Volcano, Montserrat, British West Indies, in 1996–1997 (Voight et al. 1998, 1999). At Soufrière Hills, periodicity in seismicity and tilt ranged from ~ 4 to 30 h, and the oscillations in both records continued for weeks. Cyclic behavior was first observed in the seismicity (RSAM) records beginning in July 1996, when the record of dome growth constrained the average supply rate to between 2 and 3 m^3/s (Sparks et al. 1998). The oscillations in the RSAM records initially had low amplitudes, and no tilt-measurement station was close enough to the vent to detect any pressure oscillations in the conduit. By August 1996, RSAM records showed strong oscillatory seismicity at dome growth rates between 3 and 4 m^3/s . Tilt data, taken close enough to the vent (i.e., Chances Peak (Voight et al. 1998)) to be sensitive to conduit pressure oscillations, are only available for February 1997 and May–August 1997. In the latter period, dome growth rate increases from $\sim 5 \text{ m}^3/\text{s}$ in May to between 6 and 10 m^3/s in August. Both near-vent tilt and RSAM displayed

oscillatory behavior during this period and were strongly correlated in time. Similar RSAM oscillations having periods of 7–10 h were observed at Mount Pinatubo following the climactic eruption in 1991 (Hoblitt et al. 1996). At both volcanoes, oscillation periods were observed that do not fit any multiple of Earth or ocean tides (Fig. 2).

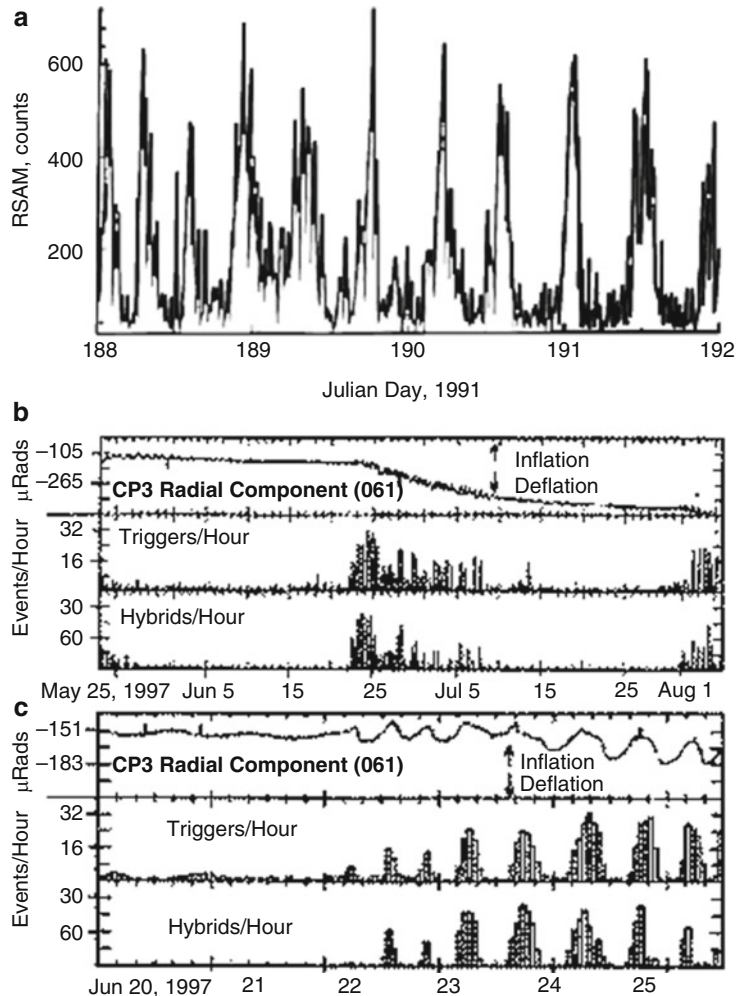
The cyclic activities at both Pinatubo and Soufrière Hills Volcano are strongly correlated with eruptive behavior and other geophysical phenomena. In the Pinatubo case and on Soufrière Hills Volcano in August, September, and October 1997, the cycles were linked to short-lived volcanic explosions. In the case of the Soufrière Hills Volcano, explosions in August 1997 occurred at the peak in the tilt. The peak in tilt also marked the

onset of episodes of increased rockfalls (Calder et al. 2002, 2005, and Voight et al. 1999) and is attributed to increased magma discharge rates. SO₂ flux data show that the cycles are linked to surges in gas release, which reach a peak about an hour after the tilt peak (Watson et al. 2000). Green and Neuberg (2006) have shown that several families of near-identical long-period earthquakes occur during the tilt cycle, starting at the inflection point on the up cycle and finishing before the inflection point on the down cycle.

Several models (Denlinger and Hoblitt 1999; Lensky et al. 2007; Neuberg et al. 2006; Wylie et al. 1999) have been proposed to explain the observed cyclicity. In these models (Denlinger and Hoblitt 1999; Wylie et al. 1999), the conduit

Volcanic Eruptions: Cyclicity During Lava Dome Growth,

Fig. 2 Examples of cyclic behavior in a lava dome eruption. RSAM records from Mt. Pinatubo, Philippines, following its climactic eruption in June 1991 ((a), after Denlinger and Hoblitt 1999). Radial tilt, triggered earthquakes, and hybrid earthquakes for 17 May to 6 August 1997, at Soufrière Hills Volcano ((b), after Costa et al. 2007b; Voight et al. 1998). Parts of three 5–7-week cycles are shown, with each cycle showing high amplitude tilt and seismicity pulsations with short timescale that last for several weeks after the start of the cycle. The onset of a cycle is rapid, as detailed in c for the cycle initiating on 22 June 1997



is divided into two parts. Magma is assumed to be forced into the lower part of volcanic conduit at a constant rate. In Denlinger and Hoblitt (1999), magma in the lower part of the conduit is assumed to be compressible. In Wylie et al. (1999), the magma is incompressible, but the cylindrical conduit is allowed to expand elastically. In both models the lower part of the conduit, therefore, acts like a capacitor that allows magma to be stored temporally in order to release it during the intense phase of the eruption. In the upper part of the conduit, friction is dependent on magma discharge rate, with a decrease in friction resulting in an increase in discharge rate, over a certain range of discharge rates. In Denlinger and Hoblitt (1999), when magma discharge rate reaches a critical value, magma detaches from the conduit walls, and a stick-slip transition occurs. Rapid motion of magma leads to depressurization of the conduit and a consequent decrease in discharge rate, until at another critical value the magma again sticks to the walls. Pressure starts to increase again due to influx of new magma into the conduit. On a pressure-discharge diagram, the path of eruption is represented by a hysteresis loop. In Wylie et al. (1999), the friction is controlled by volatile-dependent viscosity. Volatile exsolution delay is controlled by diffusion. When magma ascends rapidly, volatiles have no time to exsolve and viscosity remains low. Depressurization of the upper part of the conduit leads to a decrease in magma discharge rate and an increase in viscosity due to more intense volatile exsolution.

In Neuberg et al. (2006), a steady 2D conduit flow model was developed. The full set of Navier-Stokes equations for a compressible fluid with variable viscosity was solved by means of a finite element code. Below some critical depth, the flow was considered to be viscous and Newtonian, with a no-slip boundary condition at the wall. Above this depth a plug develops, with a wall boundary condition of frictional slip. The slip criterion was based on the assumption that the shear stress inside the magma overcomes some critical value. Simulations reveal that slip occurs in the shallow part of the conduit, in good agreement with locations of long-period volcanic earthquakes for the

Soufrière Hills Volcano (Green and Neuberg 2006). However, some parameters used in the simulations (low crystallinity, less than 30%, and high discharge rate, more than $100 \text{ m}^3/\text{s}$) are inconsistent with observations.

Lensky et al. (2007) developed the stick-slip model by incorporating degassing from supersaturated magma together with a sticking plug. Gas diffuses into the magma, which cannot expand due to the presence of a sticking plug, resulting in a buildup of pressure. Eventually the pressure exceeds the strength of the plug, which fails in stick-slip motion, and the pressure is relieved. The magma sticks again when the pressure falls below the dynamic friction value. In this model the time-scale of the cycles is controlled by gas diffusion. The influence of permeable gas loss, crystallization, and elastic expansion of the conduit on the period of pulsations was studied.

A shorter timescale of order of minutes was investigated in Iverson et al. (2006) in relation to repetitive seismic events during the 2004–2006 eruption of Mount St. Helens. The flow dynamics is controlled by the presence of a solid plug that is pushed by a Newtonian liquid, with the possibility of a stick-slip transition. Inertia of the plug becomes important on such short timescales.

Models to explain the occurrence of Vulcanian explosions have also been developed by Connor et al. (2003), Jaquet et al. (2006), and Clarke et al. (2007). A statistical model of repose periods between explosions by Connor et al. (2003) shows that data fit a log-logistic distribution, consistent with the interaction of two competing processes that decrease and increase gas pressure, respectively. Jaquet et al. (2006) show the explosion repose period data have a memory. The petrological observations of Clarke et al. (2007) on clasts from Vulcanian explosions associated with short-term cycles support a model where pressure builds up beneath a plug by gas diffusion but is opposed by gas leakage through a permeable magma foam. Although these models include some of the key processes and have promising explanatory power, they do not consider the development of the magma plug explicitly. This process is considered in Diller et al. (2006), where a model of magma ascent with gas escape is proposed.

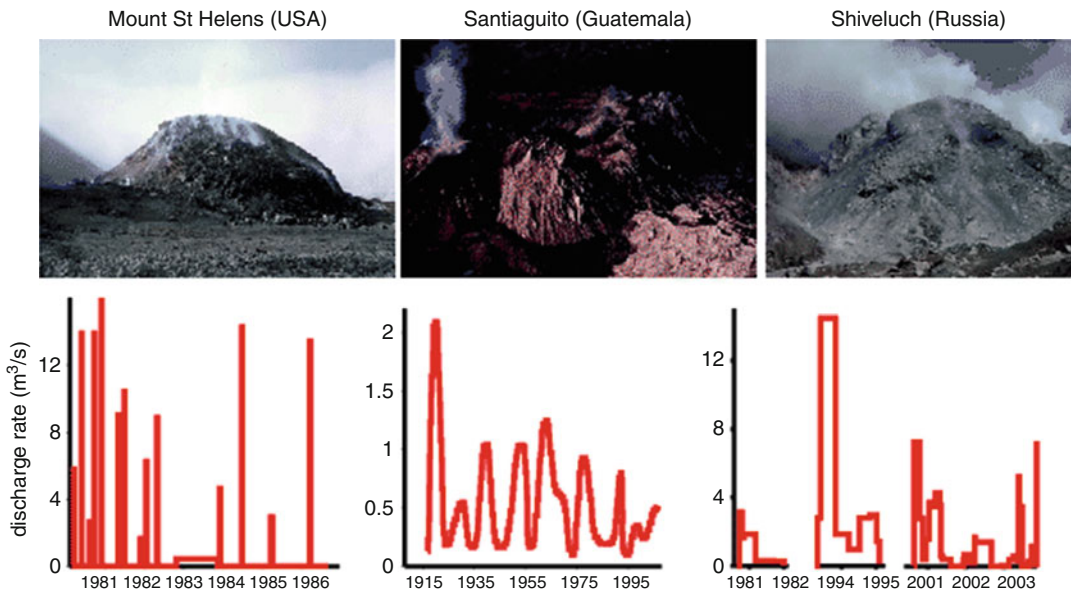
Long-Term Cycles

Figure 3 shows views of three lava domes (Mount St. Helens, USA; Santiaguito, Guatemala; and Shiveluch, Russia). Variations in magma discharge rate with time are presented below (Dirksen et al. 2006; Harris et al. 2002; Swanson and Holcomb 1990). Behavior at the first two volcanoes is rather regular, whereas at Shiveluch, long repose periods are followed by an initial rapid increase in eruptive activity with subsequent decrease and complete stop of the eruption. Growth of the lava dome at Unzen volcano, Japan, 1991–1995, was similar to Shiveluch (Dirksen et al. 2006; Nakada et al. 1999).

There are three types of conceptual models that attempt to explain long-term variation in magma discharge rate. Maeda (2000), after Ida (1996), considers a simple system that contains a spherical magma chamber located in elastic rocks with a cylindrical conduit located in viscoelastic rocks. Magma viscosity is assumed to be constant. The magma chamber is replenished with a time-dependent influx rate. The model reproduces discharge rate variation at Unzen volcano by

assuming a bell-shaped form of influx rate dependence on time. There are two controversial assumptions in the model. First, the assumption that the conduit wall rocks are viscoelastic, while the magma chamber wall rocks are purely elastic, cannot be justified because near the chamber rock temperature is the highest and it is more reasonable to expect viscous properties for chamber wall rocks rather than for the conduit. The equation that links conduit diameter with magmatic overpressure assumes viscous rock properties up to infinity. If the chamber is located in viscoelastic rocks, oscillations in discharge rate are not possible. Second, in order to obtain reasonable timescales, the rock viscosity must be rather small, of order of 10^{13} Pa s, which is only slightly higher than the typical viscosities of the magma.

Another set of models attribute cyclic behavior to heat exchange between ascending magma and wall rocks, which accounts for temperature-dependent viscosity (Costa and Macedonio 2002; Whitehead and Helfrich 1991). The idea of both models is that magma cools down as it ascends and heat flux is proportional to the



Volcanic Eruptions: Cyclicity During Lava Dome Growth, Fig. 3 Observed discharge rate versus time for (a) Mount St. Helens dome growth and (b) Santiaguito

volcano and Shiveluch (c). Photos: Mount St. Helens by Lyn Topinka (1985), Santiaguito by Gregg Bluth (2002), Shiveluch by Pavel Plechov (2001)

difference between the average temperature of the magma and the temperature of the wall rocks. If magma ascends quickly, heat loss is small in comparison with heat advection. Magma viscosity remains low as a consequence and allows high magma discharge rates. In contrast, when magma ascends slowly, it can cool substantially, and viscosity increases significantly. Both models suggest that for a fixed chamber pressure, there can be up to three steady-state solutions with markedly different discharge rates. Transition between these steady-state solutions leads to cyclic variations in discharge rate. Whitehead and Helfrich (1991) demonstrated the existence of cyclic regimes in experiments using corn syrup. In application to magma ascent in a volcanic conduit, these models have strong limitations, because a constant wall rock temperature is assumed. However, as an eruption progresses, the wall rocks heat up, and heat flux decreases, a condition that makes periodic behavior impossible for long-lived eruptions. For such a long-lived eruption like Santiaguito (started in 1922), wall rocks are expected to be nearly equilibrated in temperature with the magma, and heat losses from magma became small. It is possible that this decrease in heat flux contributes to a slow progressive increase in temperature that is observed on timescales longer than the period of pulsations. For example, magma at Santiaguito becomes progressively less viscous, resulting in a transition from mainly lava dome to lava flow activity.

Models, developed by authors of this manuscript, consider that degassing-induced crystallization is a major controlling process for the long-term cyclicity during lava dome-building eruptions. There is increasing evidence that there is a good correlation between magma discharge rate and crystallinity of the magma (Cashman and Blundy 2000; Nakada et al. 1999). An increase in crystal content leads to an increase in magma viscosity (Costa 2005; Costa et al. 2007a,b) and, thus, influences magma ascent dynamics. First, we consider a simplified model of magma ascent in a volcanic conduit that accounts for crystallization and rheological stiffening.

A Simplified Model

In Barmin et al. (2002), the following simplifying assumptions have been made in order to develop a semi-analytical approach to magma ascent dynamics.

1. Magma is incompressible. The density change due to bubble formation and melt crystallization is neglected.
2. Magma is a viscous Newtonian fluid. Viscosity is a step function of crystal content. When the concentration of crystals β reaches a critical value β_* , the viscosity of magma increases from value μ_1 to a higher value μ_2 . Later on we will consider magma rheology in more detail (see section “[Rheology of Crystal-Bearing Magma and Conduit Resistance](#)”), but a sharp increase in viscosity over a narrow range of crystal content has been confirmed experimentally (e.g., Caricchi et al. 2007; Lejeune and Richet 1995).
3. Crystal growth rate is constant and no nucleation occurs in the conduit. The model neglects the fact that magma is a complicated multi-component system, and its crystallization is controlled by the degree of undercooling (defined as the difference between actual temperature of the magma and its liquidus temperature). Later in the entry, a more elaborate model for magma crystallization will be considered.
4. The conduit is a vertical cylindrical pipe. Elastic deformation of the wall rocks is not included in the model. This assumption is valid for a cylindrical shape of the conduit at typical magmatic overpressures but is violated when the conduit has a fracture shape. Real geometries of volcanic conduits and their inclination can vary significantly with depth.
5. The magma chamber is located in elastic rocks and is fed from below, with a constant influx rate. For some volcanoes, like Santiaguito or Mount St. Helens, average magma discharge rate remained approximately constant during several periods of pulsation. Thus the assumption of constant influx rate is valid. For volcanoes like Mount Unzen or Shiveluch, there is

an evidence of pulse-like magma recharge (Dirksen et al. 2006; Maeda 2000).

With above simplification the system of equations for unsteady 1D flow is as follows:

$$\frac{\partial}{\partial t}\rho + \frac{\partial}{\partial x}\rho u = 0 \;; \quad \frac{\partial}{\partial t}n + \frac{\partial}{\partial x}nu = 0 \quad (1a)$$

$$\frac{\partial p}{\partial x} = -\rho g - \frac{32\mu u}{\delta^2} \;; \quad \mu = \begin{cases} \mu_1, & \beta < \beta_* \\ \mu_2, & \beta \geq \beta_* \end{cases} \quad (1b)$$

$$\frac{\partial}{\partial t}\beta + u\frac{\partial\beta}{\partial x} = 4\pi nr^2\chi = (36\pi n)^{\frac{1}{3}}\beta^{\frac{2}{3}}\chi \quad (1c)$$

Here ρ is the density of magma, u is the vertical cross-sectional averaged ascent velocity, n is the number density of crystals per unit volume, p is the pressure, g is the acceleration due to gravity, δ is the conduit diameter, β is the volume concentration of crystals, β_* is a critical concentration of crystals above which the viscosity changes from μ_1 to μ_2 , r is the crystal radii, χ is the linear crystal growth rate, and x is the vertical coordinate. The first two Eqs. (1a) represent the conservation of mass and the number density of crystals, the second (1b) is the momentum equation with negligible inertia, and the third (1c) is the crystal growth equations with $\chi = \text{constant}$. We assume the following boundary conditions for the system (1a–1c):

$$x = 0 : \frac{dp_{\text{ch}}}{dt} = \frac{\gamma}{V_{\text{ch}}}(Q_{\text{in}} - Q_{\text{out}}); \beta = \beta_{\text{ch}}; n = n_{\text{ch}}; x = l : p = 0$$

Here γ is the rigidity of the wall rock of the magma chamber, V_{ch} is the chamber volume, β_{ch} and n_{ch} are the crystal concentration and number density of crystals per unit volume in the chamber, p_{ch} is the pressure in the chamber, l is the conduit length, Q_{in} is the flux into the chamber, and $Q_{\text{out}} = \pi\delta^2u/4$ is the flux out of the chamber into the conduit. Both β_{ch} and n_{ch} are assumed constant. We neglect the influence of variations of the height of the lava dome on the pressure at the top of the conduit and assume that the pressure there is constant. As the magma is assumed to be incompressible, the pressure at the top of the conduit can

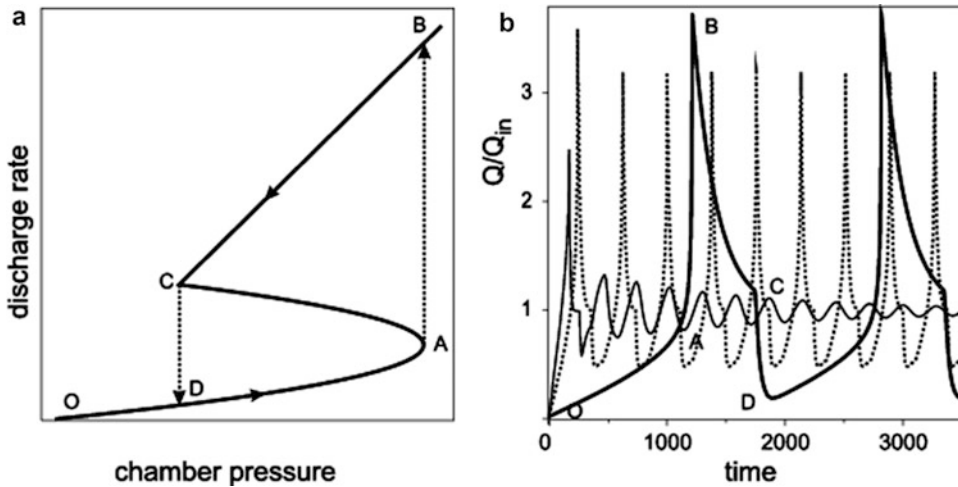
be set to zero because the atmospheric pressure is much smaller than magma chamber pressure.

From the mass conservation equation for the case of constant magma density, $u = u(t)$ and $n = n_{\text{ch}}$ everywhere. Equations (1) can be integrated and transformed from partial differential equations to a set of ordinary differential equations with state-dependent delay representing a “memory” effect on crystal concentration (see Barmin et al. 2002 for details).

Results and Applications

The general steady-state solution for magma ascent velocity variations with chamber pressure is shown in Fig. 4a. Solutions at high magma ascent velocities, when the critical concentration of crystals is not reached inside the conduit, result in a straight line (CB), which is the same as for the classical Poiseuille solution for a fluid with constant viscosity. At low ascent velocities, there is a quadratic relationship (OAC) between chamber pressure and ascent velocity (see Barmin et al. 2002 for derivation of the equation). A key feature of the steady-state solution is that, for a fixed chamber pressure, it is possible to have three different magma ascent velocities. We note that for $\mu_2/\mu_1 = 1$ only the branch CB exists and for $1 < \mu_2/\mu_1 \leq 2$ there is a smooth transition between the lower branch (ODA) and the upper branch (CB) and multiple steady-state regimes do not exist.

We first consider the case where chamber pressure changes quasi-statically, and the value of Q_{in} is between Q_A and Q_C . Starting at point O, the chamber pressure increases, because the influx into the chamber is higher than the outflux. At point A, a further increase in pressure is not possible along the same branch of the steady-state solution, and the system must change to point B, where the outflux of magma is larger than the influx. The chamber pressure and ascent velocity decrease along BC until the point C is reached and the system must change to point D. The cycle DABC then repeats itself. Provided the chamber continues to be supplied at the same constant rate repetition of this cycle results in periodic behavior. The transitions AB and CD in a cycle must involve unsteady flow.



Volcanic Eruptions: Cyclicity During Lava Dome Growth, Fig. 4 The general steady-state solution and possible quasi-static evolution of an eruption. After Barmin et al. (2002). Different curves correspond to

different values of dimensionless parameter $\kappa = (\pi\delta^2\gamma)/(4V_{ch}\rho g)$ (0.12, thin line; 0.05, dotted line; and 0.005, bold line)

Oscillations in magma discharge rate involve large variations in magma crystal content. This relation is observed on many volcanoes. For example, pumice and samples of the Soufrière Hills dome that were erupted during periods of high discharge (Sparks et al. 1998) have high glass contents (25% to 35%) and few microlites (Murphy et al. 2000), whereas samples derived from parts of the dome that were extruded more slowly (days to weeks typically) have much lower glass contents (5% to 15%) and high contents of groundmass microlites. These and other observations (Hammer and Rutherford 2002; Nakada et al. 1999) suggest that microlite crystallization can take place on similar timescales to the ascent time of the magma.

Of more general interest is to consider unsteady flow behavior. We assume that the initial distribution of parameters in the conduit corresponds to the steady-state solution of system (1) with the initial magma discharge rate, Q_0 , being in the lowest regime. The behavior of an eruption with time depends strongly on the value of Q_{in} . If Q_{in} corresponds to the upper or the lower branch of the steady-state solution, the eruption stabilizes with time with $Q = Q_{in}$ and $dp_{ch}/dt = 0$. However, if Q_{in} corresponds to the intermediate

branch of the steady-state solution, with Q_{in} between Q_A and Q_C , periodic behavior is possible. Figure 4b shows three eruption scenarios for different values of the magma chamber volume V_{ch} . When V_{ch} is small, the eruption stabilizes with time. In contrast, undamped periodic oscillations occur for values of V_{ch} larger than some critical value. For very large magma chamber volumes, the transient solution almost exactly follows the steady-state solution, with unsteady transitions between the regimes. The time that the system spends in unsteady transitions in this case is much shorter than the period of pulsations. The maximum discharge rate during the cycle is close to Q_B , the minimum is close to Q_D , and the average is equal to Q_{in} . The period of pulsations increases as the volume of magma chamber increases.

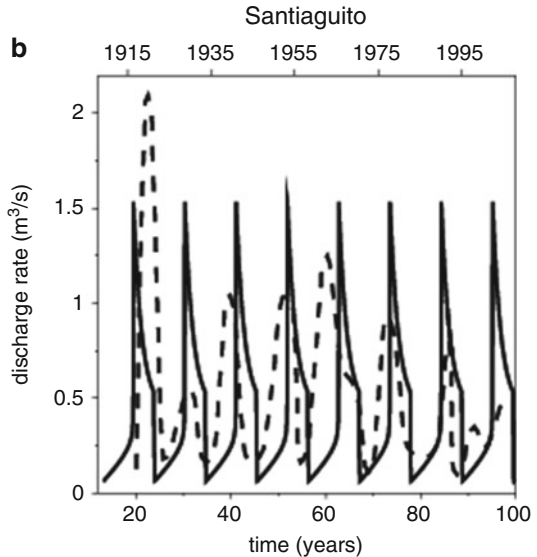
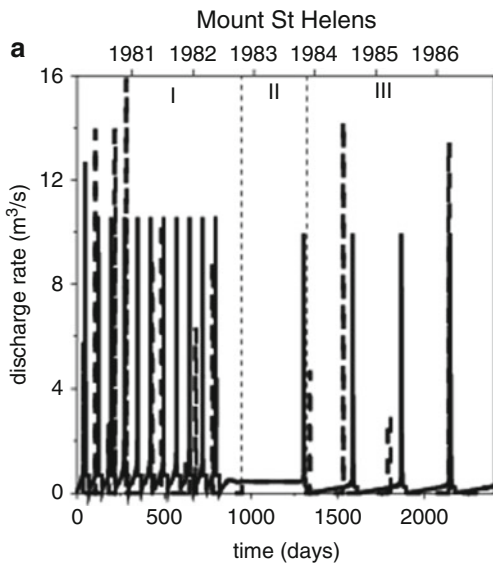
Now we apply the model to two well-documented eruptions: the growth of lava domes on Mount St. Helens (1980–1986) and on Santiaguito (1922–present). Our objective here is to establish that the model can reproduce the periodic behaviors observed at these two volcanoes. Estimates can be obtained for most of the system parameters. However, magma chamber size is not well-constrained, and so the model can be used to

make qualitative inferences on relative chamber size. Given the uncertainties in the parameter values and the simplifications in the model development, the approach can be characterized as mimicry. Adjustments in some parameters were made to achieve best fits with observations, but the particular best fits are not unique.

For Mount St. Helens, our model is based on data presented by Swanson and Holcomb (1990). Three periods of activity can be distinguished during the period of dome growth. The first period consists of 9 pulses of activity with average peak magma discharge rates $\sim 15 \text{ m}^3/\text{s}$ during 1981–1982. Each pulse lasted from 2 to 7 days (with a mean value of 4 days) with the average period between the pulses being 74 days and the average discharge rate during the period being $Q_1 \sim 0.67 \text{ m}^3/\text{s}$. The second period is represented by continuous dome growth and lasted more than a year (368 days) with a mean magma discharge rate of $Q_2 \sim 0.48 \text{ m}^3/\text{s}$. During the last period, there were 5 episodes of dome growth with peak magma discharge rates up to $\sim 15 \text{ m}^3/\text{s}$, an average period of pulsation of ~ 230 days, and a mean

discharge rate of $Q_3 \sim 0.23 \text{ m}^3/\text{s}$. We assume that the intensity of influx into the magma chamber, Q_{in} , is equal to the average magma discharge rate over the corresponding periods. There might have been a progressive decrease in the intensity of influx during the eruption, but, due to limitations of the model, we assume that Q_{in} changes as a step function between the periods.

The best-fit model for the eruption is presented in Fig. 5a, with the parameters used for the simulation summarized in Table 2 in Barmin et al. (2002). During the first period of the eruption, $Q_{\text{in}} = Q_1$. This corresponds to the intermediate branch of the steady-state solution, and cyclic behavior occurs. In the second period $Q_{\text{in}} = Q_2$, the system moves to the lower regime, and the eruption stabilizes with time. For the third period of the eruption, the parameters of the system have been changed, so that $Q_{\text{in}} = Q_3 < Q_2$, corresponding to the intermediate regime once again, and periodic behavior occurs. This condition can be satisfied by a decrease in the diameter of the conduit, a decrease in crystal growth rate, or the number



Volcanic Eruptions: Cyclicity During Lava Dome Growth, Fig. 5 Discharge rate versus time for (a) Mount St. Helens dome growth and (b) Santiaguito

volcano. Dotted lines represented are the observed values of discharge rates, and the solid lines are the best-fit simulations. (After Barmin et al. 2002)

density of crystals. All these mechanisms are possible: decrease in the diameter could be a consequence of magma crystallization on the conduit walls, while a decrease in either crystal growth rate or number density of crystals could reflect observed changes in magma composition (Swanson and Holcomb 1990). The influence of conduit diameter is the strongest because ascent velocity, for the same discharge rate, depends on the square of the diameter. The required change in diameter is from 18 to 12 m, but this change can be smaller if we assume a simultaneous decrease in crystal growth rate.

Since 1922, lava extrusion at Santiaguito has been cyclic (Harris et al. 2002). Each cycle begins with a 3–6 year long high ($0.5\text{--}2.1\text{ m}^3/\text{s}$) magma discharge rate phase, followed by a longer (3–11 years) low ($\sim 0.2\text{ m}^3/\text{s}$) discharge rate phase. The time-averaged magma discharge rate was almost constant at $\sim 0.44\text{ m}^3/\text{s}$ between 1922 and 2000. The first peak in discharge rate had a value $>2\text{ m}^3/\text{s}$, whereas the second peak had a much smaller value. The value for the second peak is underestimated as it is calculated based on the dome volume only but does not include the volume of dome-collapse pyroclastic flows. Later peaks show an increase in magma discharge rate until 1960 (Fig. 5b, dashed line). Post-1960, the duration of the low discharge rate phase increased, the peak discharge and the time-averaged discharge rates for each cycle decreased, and the difference between discharge rates during the high and low discharge rate phases of each cycle decreased. Our best-fit model is shown in Fig. 5b, and the parameter estimates are listed in Table 2 in Barmin et al. (2002). The model reproduces the main features of the eruption, including the period of pulsations, the ratio between low and high magma discharge rates, and the range of observed discharge rates. We cannot, however, reproduce the decrease in the amplitude of pulsations within the framework of the model using fixed parameter values.

The theory provides a potential method to estimate magma chamber volumes. For Mount St. Helens, our estimate of the chamber size (0.6 km^3) is comparable with the total erupted

volume in the entire 1980–1986 eruption and is consistent with the fact that geophysical imaging did not identify a large magma body. Santiaguito volcano erupted more than 10 km^3 in the 1902 explosive eruption (Williams and Self 1983) and more than 1 km^3 of lava domes since 1922. The best-fit model estimate of a large (64 km^3) chamber is consistent with much larger eruption volumes, long periods, and longevity of the eruption in comparison to Mount St. Helens. One limitation of the model is that the supply of deep magma from depth to the chamber is assumed to be constant.

Model Development

In this section, we further develop models to examine new effects and relax some of the simplifications of earlier models. We investigate a number of effects that were not fully explained or considered in previous studies (Barmin et al. 2002; Melnik and Sparks 1999, 2002). The new model incorporates a more advanced treatment of crystallization kinetics based on the theoretical concepts developed in (Hort 1998; Kirkpatrick 1976) and is calibrated by experimental studies in andesitic systems (Couch et al. 2001; Hammer and Rutherford 2002). In particular, we distinguish growth of phenocrysts formed in the magma chamber from crystallization of microlites during magma ascent. Previous models have assumed that magma is always Newtonian, so we study models of conduit flow assuming non-Newtonian rheology, with rheological properties being related to crystal content. Latent heat is released during the crystallization of ascending magma due to degassing, and we show that this can have an important influence on the dynamics. Elastic deformation of conduit walls leads to coupling between magma ascent and volcano deformations.

System of Equations

We model magma ascent in a dyke-shaped conduit with elliptical cross section using a set of 1D transient equations written for horizontally averaged variables (Costa et al. 2007a,b):

$$\frac{1}{S} \frac{\partial}{\partial t} (S\rho_m) + \frac{1}{S} \frac{\partial}{\partial x} (S\rho_m V) = -G_{mc} - G_{ph} \quad (2)$$

$$\frac{1}{S} \frac{\partial}{\partial t} (S\rho_{mc}) + \frac{1}{S} \frac{\partial}{\partial x} (S\rho_{mc} V) = G_{mc} \quad (3a)$$

$$\frac{1}{S} \frac{\partial}{\partial t} (S\rho_{ph}) + \frac{1}{S} \frac{\partial}{\partial x} (S\rho_{ph} V) = G_{ph} \quad (3b)$$

$$\frac{1}{S} \frac{\partial}{\partial t} (S\rho_d) + \frac{1}{S} \frac{\partial}{\partial x} (S\rho_d V) = -J \quad (4a)$$

$$\frac{1}{S} \frac{\partial}{\partial t} (S\rho_g) + \frac{1}{S} \frac{\partial}{\partial x} (S\rho_g V_g) = J \quad (4b)$$

Here t denotes time; x is the vertical coordinate; ρ_m , ρ_{ph} , ρ_{mc} , ρ_d , and ρ_g are the densities of melt, phenocrysts, microlites, dissolved gas, and exsolved gas, respectively; and V and V_g are the velocities of magma and gas, respectively. G_{ph} and G_{mc} represent the mass transfer rate due to crystallization of phenocrysts and microlites, respectively, J is the mass transfer rate due to gas exsolution, and S is the cross-sectional area of the conduit. Equation (2) represents the mass conservation for the melt phase; Eqs. (3a) and (3b) are the conservation equations for microlites and phenocrysts, respectively; and Eqs. (4a) and (4b) represent the conservation of the dissolved gas and of the exsolved gas, respectively.

$$\frac{\partial p}{\partial x} = -\rho g - F_c \quad (5)$$

$$V_g - V = -\frac{k}{\mu_g} \frac{\partial p}{\partial x} \quad (6)$$

Here p is the pressure, ρ is the bulk density of magma, g is the acceleration due to gravity, μ is the magma viscosity, k is the magma permeability, and μ_g is the gas viscosity. Equation (5) represents the equation of momentum for the mixture as a whole, in which the pressure drops due to gravity and conduit resistance are calculated for laminar flow in an elliptic pipe. Equation (6) is the Darcy law for the exsolved gas flux through the magma.

$$\begin{aligned} \frac{1}{S} \frac{\partial}{\partial t} (S\rho C_m T) + \frac{1}{S} \frac{\partial}{\partial x} (S\rho C_m VT) \\ = L_* (G_{mc} + G_{ph}) - C_m TJ - Q_{cl} + Q_{vh} \end{aligned} \quad (7)$$

Here C_m is the bulk-specific heat of magma, T is the bulk flow-averaged temperature, L_* is latent heat of crystallization, Q_{cl} denotes the total heat loss by conduction to the conduit walls, and Q_{vh} denotes the total heat generation due to viscous dissipation. Here we consider the case of the latent heat release only. This assumption is valid when both $Q_{cl} \approx 0$ and $Q_{vh} \approx 0$ or when $Q_{cl} + Q_{vh} \approx 0$. The study of the effects of both heat loss and viscous heating, which are intrinsically two-dimensional (Costa and Macedonio 2003, 2005), and their parametrization is the subject of ongoing research.

$$\begin{aligned} \rho_m &= \rho_m^0 (1 - \alpha)(1 - \beta)(1 - c); \quad \rho_c \\ &= \rho_c^0 (1 - \alpha)\beta \end{aligned} \quad (8a)$$

$$\rho_d = \rho_m^0 (1 - \alpha)(1 - \beta)c; \quad \rho_g = \rho_g^0 \alpha \quad (8b)$$

$$\rho = \rho_m + \rho_c + \rho_d + \rho_g \quad (8c)$$

$$\begin{aligned} \alpha &= \frac{4}{3} \pi r_b^3 n; \quad \frac{\partial}{\partial t} (Sn) + \frac{\partial}{\partial x} (SnV) = 0; \\ p &= \rho_g^0 RT \end{aligned} \quad (9)$$

Here α is the volume concentration of bubble, β is the volume concentration of crystals in the condensed phase (melt plus crystals), c is mass concentration of dissolved gas (equal to volume concentration as we assume that the density of dissolved volatiles is the same as the density of the melt), ρ_m^0 denotes the mean density of the pure melt phase, ρ_c^0 is density of the pure crystal phase (with $\rho_c = \rho_{ph} + \rho_{mc}$, $\beta = \beta_{ph} + \beta_{mc}$), r_b is the bubble radius, and n is the number density of bubble per unit volume. Concerning the parametrization of mass transfer rate functions, we use

$$J = 4\pi r_b n D \rho_m^0 (c - C_f \sqrt{p}) \quad (10)$$

$$G_{mc} = 4\pi\rho_c^0(1-\beta)(1-\alpha) \times U(t) \int_0^t I(\omega) \left(\int_\omega^t U(\eta) d\eta \right)^2 d\omega \quad (11a)$$

$$G_{ph} = 3\gamma_s \left(\frac{4\pi N_{ph}\beta_{ph}^2}{3} \right)^{\frac{1}{3}} \rho_c^0(1-\beta)(1-\alpha)U(t) \quad (11b)$$

Here J is parametrized using the analytical solution described in Navon and Lyakhovsky (1998); U is the linear crystal growth rate (ms^{-1}); I is the nucleation rate ($m^{-3}s^{-1}$), which defines the number of newly nucleated crystal per cubic meter; γ_s is a shape factor of the order of unity; and D and C_f are the diffusion and the solubility coefficients, respectively. Concerning the mass transfer due to crystallization G_{mc} , we adapt a model similar to that described in Hort (1998). Assuming spherical crystals, the Avrami-Johnson-Mehl-Kolmogorov equation in the form adopted by Kirkpatrick (1976), for the crystal volume increase rate, is

$$\frac{d\beta}{dt} = 4\pi Y_t U(t) \int_0^t I(\omega) \left(\int_\omega^t U(\eta) d\eta \right)^2 d\omega$$

where $Y_t = (1-\beta)(1-\alpha)$ is the volume fraction of melt remaining uncrystallized at the time t . Therefore, we have $G_{mc} = \rho_{mc} d\beta/dt$. For the phenocryst growth rate $G_{ph}(0)$, we assume that it is proportional to the phenocryst volume increase rate $d\beta_{ph}/dt = 4\pi R_{ph}^2 N_{ph} U(t)$ times the crystal density ρ_c^0 times the volume fraction of melt remaining uncrystallized at the time t . A detailed description of the parametrization used for the different terms is reported in Melnik and Sparks (2005).

For parametrizations of magma permeability k and magma viscosity μ , we use

$$k = k(\alpha) = k_0 \alpha^j \quad (12)$$

$$\mu = \mu_m(c, T) \theta(\beta) \eta(\alpha, Ca) \quad (13)$$

where k is assumed to depend only on bubble volume fraction α . Magma viscosity μ depends

on water content, temperature, crystal content, bubble fraction, and capillary number as described in detail in the next section.

Regarding equations for semiaxes, a and b , we assume that the elliptical shape is maintained and that pressure changes gradually in respect with vertical coordinate and time so that the plain strain analytical solution for an ellipse subjected to a constant internal overpressure (Mériaux and Jaupart 1995; Muskhelishvili 1963) remains valid:

$$a = a_0 + \frac{\Delta P}{2G} [-(1-2\nu)a_0 + 2(1-\nu)b_0] \quad (14a)$$

$$b = b_0 + \frac{\Delta P}{2G} [2(1-\nu)a_0 - (1-2\nu)b_0] \quad (14b)$$

where ΔP is the overpressure, i.e., the difference between conduit pressure and far-field pressure (here assumed to be lithostatic for a sake of simplicity); a_0 and b_0 are the initial values of the semiaxes; ν is the host rock Poisson ratio; and G is the host rock rigidity.

Equations (2)–(14a, 14b) are solved between the top of the magma chamber and the bottom of the lava dome that provides some constant load by using the numerical method described in Melnik and Sparks (2005). The effects of dome height and morphology changes are not considered in this entry. We consider three different kinds of boundary conditions at the inlet of the dyke: constant pressure, constant influx rate, and the presence of a magma chamber located in elastic rocks. The case of constant pressure is applicable when a dyke starts from either a large magma chamber or unspecified source, so that pressure variations in the source region remain small. An estimate of the volume of magma stored in the source region that allows pressure to be approximated as constant depends on wall rock elasticity, magma compressibility (volatile content), and the total volume of the erupted material. If the magma flow at depth is controlled by regional tectonics, the case of constant influx rate into the dyke may

be applicable if total variations in supply rate are relatively small on the timescale of the eruption.

For the case where magma is stored in a shallow magma chamber prior to eruption and significant chamber replenishment occurs, the flow inside the conduit must be coupled with the model for the magma chamber. In this case, as explained in detail in Melnik and Sparks (2005), we assume that the relationship between the pressure at the top of the magma chamber p_{ch} and the intensity of influx Q_{in} and outflux Q_{out} of magma to and from the chamber is given by

$$\frac{dp_{\text{ch}}}{dt} = \frac{4G\langle K \rangle}{\langle \rho \rangle V_{\text{ch}}(3\langle K \rangle + 4G)}(Q_{\text{in}} - Q_{\text{out}}) \quad (15)$$

where V_{ch} is the magma chamber volume; $\langle \rho \rangle$ and $\langle K \rangle$ are the average magma density and magma bulk modulus, respectively; and G is the rigidity of rocks surrounding the chamber.

Cases of constant influx and of constant source pressure are the limit cases of Eq. (15) in the case of infinitely small and infinitely large magma chamber volume. We assume that the volume concentration of bubbles and phenocrysts is determined by equilibrium conditions and that the temperature of the magma is constant. The effect of temperature change on eruption dynamics, due to interaction between silicic and basaltic magma, was studied in Melnik and Sparks (2005).

We use a steady-state distribution of parameters along the conduit as an initial condition for the transient simulation. The values are calculated for a low magma discharge rate, but the particular value of this parameter is not important because the system deviates from initial conditions to a cyclic or stabilized state, which does not depend on the initial conditions.

Rheology of Crystal-Bearing Magma and Conduit Resistance

Magma viscosity is modeled as a product of melt viscosity $\mu_{\text{m}}(c, T)$, the relative viscosity due to crystal content $\theta(\beta) = \Theta(\beta)\varphi(\beta)$, and the relative viscosity due to the presence of bubbles $\eta(\alpha, Ca)$. The viscosity of the pure melt $\mu_{\text{m}}(c, T)$ is calculated according to Hess and

Dingwell (1996). Viscosity increase due to the presence of the crystals is described through the function $\Theta(\beta)$ (Costa 2005; Costa et al. 2007a,b). As crystallization proceeds, the remaining melt becomes enriched in silica, and melt viscosity increases. The parametrization of this effect is described by the function $\varphi(\beta)$ in Costa et al. (2007b) and Dirksen et al. (2006). Effects of the solid fraction are parametrized as described in Costa et al. (2007b).

Effects due to the presence of bubbles are accounted for by adopting a generalization of Llewellyn and Manga (2005) for an elliptical conduit (Costa et al. 2007a,b).

In the case of Newtonian magma rheology, the friction force in an elliptical conduit can be obtained from a classical Poiseuille solution for low Reynolds number flow $F_c = 4\mu(a^2 + b^2)/(a^2b^2)V$ (Landau and Lifshitz 1987). High crystal or bubble content magmas may show non-Newtonian rheology. One possible non-Newtonian rheology is that of a Bingham material characterized by a yield strength τ_b (Bingham 1922). The stress-strain relation for this material is given by

$$\begin{aligned} \tau_{ij} &= \left(\mu + \frac{\tau_b}{\gamma_b} \right) \gamma_{ij} \\ &\Leftrightarrow \tau > \tau_b \gamma_{ij} = 0 \\ &\Leftrightarrow \tau \leq \tau_b \end{aligned} \quad (16)$$

Here τ_{ij} and γ_{ij} are the stress and strain rate tensors, and τ and γ_b are second invariants of corresponding tensors. According to this rheological law, the material behaves linearly when the applied stress is higher than a yield strength. No motion occurs if the stress is lower than a yield strength. In the case of a cylindrical conduit, the average velocity can be calculated in terms of the stress on the conduit wall τ_w (Loitsyansky 1978):

$$V = \frac{1}{12} \frac{r}{\tau_w^3 \mu} (\tau_b^4 + 3\tau_w^4 - 4\tau_b \tau_w^3) \quad (17)$$

Here r is the conduit radii. This form of equation gives an implicit relation between ascent velocity and pressure drop and is not convenient to use. By introducing dimensionless variables

$\Pi = \mu V/\tau_b r$ and $\Omega = \tau_w/\tau_b \geq 1$ relation (17) can be transformed into

$$\Omega^4 - \frac{1}{6}(8 + 3\Pi)\Omega^3 + \frac{1}{3} = 0 \quad (18)$$

Following Melnik and Sparks (2005), a semi-analytical solution can be used for (18), and the conduit friction force can be expressed finally as follows:

$$F_c = \frac{2\tau_w}{r} = \frac{2\tau_b\Omega(\Pi)}{r}.$$

We note that a finite pressure gradient is necessary to initiate the flow in the case of Bingham liquid, in contrast to a Newtonian liquid.

Results and Applications

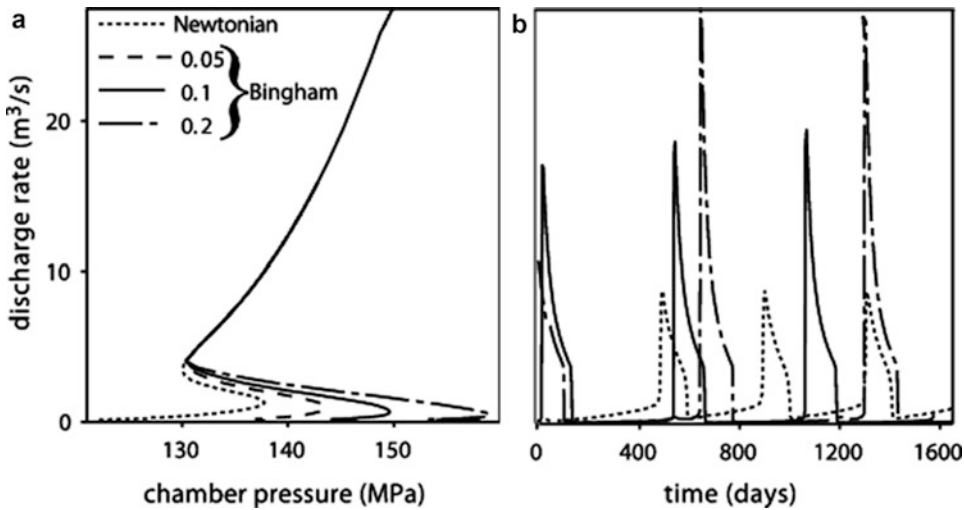
Influence of Non-Newtonian Properties on Eruption Behavior Now we compare the dynamics of magma extrusion in the cases of Newtonian and Bingham rheology. We will

assume that yield strength is reached when the concentration of crystals reaches a critical value:

$$\tau = \begin{cases} \tau_b & \text{for } \beta > \beta_{cr} \\ 0 & \text{for } \beta \leq \beta_{cr} \end{cases} \quad (19)$$

Figure 6a shows a set of steady-state solutions for different values of τ_b . Values of τ_b and β_{cr} depend on crystal shape, crystal size distribution, magma temperature, and other properties but here are assumed to be constant. To illustrate the influence of Bingham rheology, the value of $\beta_{cr} = 0.65$ was chosen so that, for discharge rate larger than $\sim 5 \text{ m}^3/\text{s}$, the magma has Newtonian rheology (see Fig. 6a). A more detailed study would require measurements of the rheological properties of magma for a wide range of crystal content and crystal size distributions. As the value of τ_b , the chamber pressure that is necessary to start the eruption increases.

Figure 6b shows the influence of these two rheological models on the dynamics of magma extrusion. In the case of Bingham rheology, magma discharge rate between the two pulses is



Volcanic Eruptions: Cyclicity During Lava Dome Growth, Fig. 6 (a) Steady-state solutions and dependence of discharge rate on time for Newtonian and Bingham rheology of the magma. Yield strength is a parameter marked on the curves (values in MPa). For Bingham rheology discharge rate remains zero between the pulses of

activity. Bingham rheology results in much higher chamber pressures prior to the onset of activity and, therefore, much higher discharge rates in comparison with Newtonian rheology. (b) Comparison of the period of pulsation in discharge rate for Newtonian and Bingham rheologies. (After Melnik and Sparks 2005)

zero until a critical chamber overpressure is reached. Then the magma discharge rate increases rapidly with decrease in crystal content, leading to a significant reduction of both magma viscosity and the length of the part of the conduit that is occupied by the Bingham liquid where $\beta_c > \beta_{cr}$. There is a transition in the system to the uppermost flow regime, and the pressure then decreases quickly. Because the pressure at the onset of the pulse was significantly larger than in the case of a Newtonian liquid, the resulting discharge rate in the case of Bingham rheology is also significantly higher.

Modeling of Conduit Flow During Dome Extrusion on Shiveluch Volcano The maximum intensity of extrusion was reached at an early stage in all three eruptions (Fig. 3). We therefore suggest that dome extrusion was initiated by high overpressure in the magma chamber with respect to the lithostatic pressure. Depressurization of the magma chamber occurred as a result of extrusion. Without magma chamber replenishment, depressurization results in a decrease in magma discharge rate. For open-system chambers, replenishment of the chamber during eruption can lead to pulsatory behavior (Barmin et al. 2002). The following account is derived from Dirksen et al. (2006). For the 1980–1981 eruption, the monotonic decrease in discharge rate indicates that there was little or no replenishment of the magma chamber. During the 1993–1995 and 2001–2004 episodes, however, the magma discharge rate fluctuated markedly, suggesting that replenishment was occurring. The influx of new magma causes an increase in magma chamber pressure and a subsequent increase in magma discharge rate. During the 2001–2004 eruption, there were at least three peaks in discharge rate. Replenishment of the magma chamber with new hot magma can explain the transition from lava dome extrusion to viscous lava flow that occurred on Shiveluch after 10 May 2004 and which continues at the time of writing (2007).

We simulated dome growth during the 2001–2002, because this dataset is the most complete and is supported by petrological

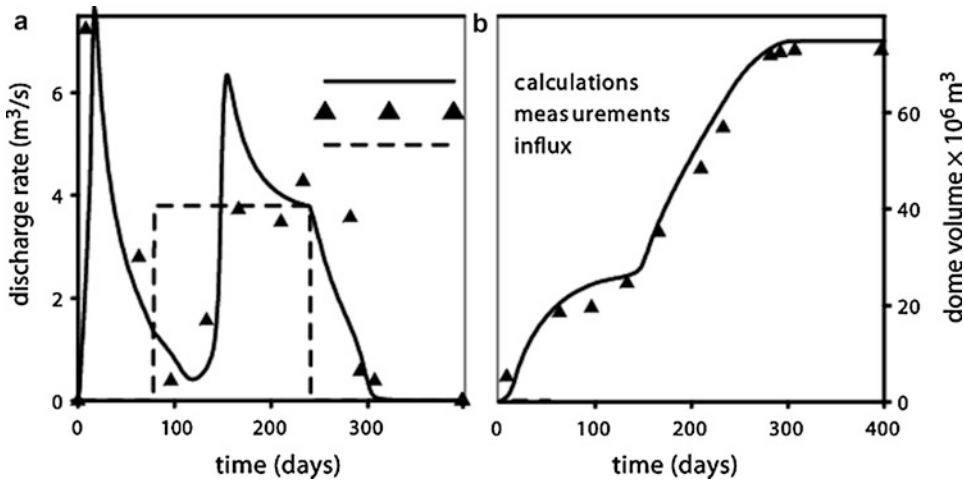
investigations (Humphreys et al. 2006). We assume the shape of the influx curve:

$$Q_{in} = \begin{cases} 0, & t < t_s \\ Q_0, & t_s \leq t \leq t_f \\ 0, & t > t_f \end{cases} \quad (20)$$

Influx occurs with constant intensity Q_0 between times t_s and t_f . We have examined many combinations of values of these parameters within the constraints provided by observations. The best simulation results use the following values of parameters: $Q_0 = 3.8 \text{ m}^3/\text{s}$, $t_s = 77$, and $t_f = 240$ days. A more continuous influx, dependent on time, is plausible, but there is no geophysical evidence that allows us to constrain the intensity of the influx, because ground deformation data are absent for Shiveluch volcano. The output of the model gives a magma chamber volume of 12 km^3 , assuming a spherical chamber. Figure 7a shows the time dependence of magma discharge rate, and Fig. 7b shows the increase in the volume of erupted material with time after 6 June 2001. The timing of magma influx is in good agreement with the residence time of basaltic magma in the system, as calculated from the olivine reaction rims. For further details see Dirksen et al. (2006).

Five- to Seven-Week Cycles on the Soufriere Hills Volcano: Evidence for a Dyke? An approximately 5- to 7-week cyclic pattern of activity was recognized at the Soufrière Hills Volcano (SHV) (Sparks and Young 2002; Voight et al. 1999) between April 1997 and March 1998 from peaks in the intensity of eruptive activity and geophysical data, including tilt and seismicity (Fig. 2).

In models discussed above, the timescale of pulsations depends principally on the volume of the magma chamber, magma rheology, and the cross-sectional area of the conduit. These models might provide an explanation for the 2–3-year cycles of dome extrusion observed at SHV, where deformation data indicate that the magma chamber regulates the cycles. However, the models cannot simultaneously explain the



Volcanic Eruptions: Cyclicity During Lava Dome Growth, Fig. 7 (a) Comparison of calculated and measured discharge rates (a) and volumes of the dome (b) for

the episode of the dome growth in 2001–2002. Influx into the magma chamber is shown by a *dashed line* in (a). Time in days begins on 6 June 2001. (After Dirksen et al. 2006)

5–7-week cycles. Thus another mechanism is needed.

The evidence for a dyke feeder at SHV includes GPS data (Mattioli et al. 1998), distribution of active vents, and seismic data (Roman et al. 2006). We have assumed that, at depth, the conduit has an elliptical shape that transforms to a cylinder at shallow level. In order to get a smooth transition from the dyke at depth to a cylindrical conduit (see Fig. 1), the value of a_0 in Eq. (14) is parametrized as follows:

$$a_0(x) = A_1 \arctan\left(\frac{x - L_T}{w_T}\right) + A_2 \quad (21)$$

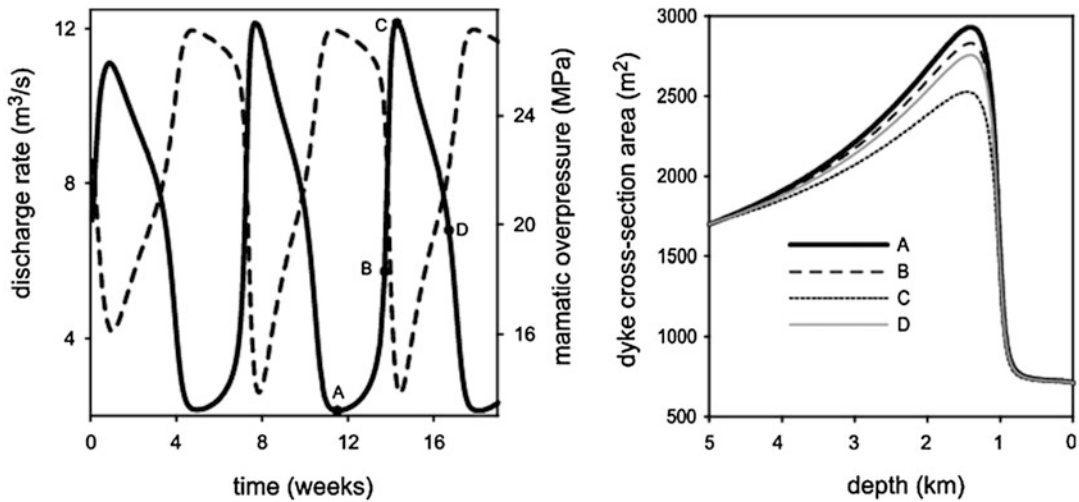
Here L_T and w_T are the position and the vertical extent of the transition zone between the ellipse and the cylinder and constants A_1 and A_2 are calculated to satisfy conditions $a_0(L) = R$ and $a_0(0) = a_0$, where R is the radius of the cylindrical part of the conduit and a_0 is the length of semi-major axis at the inlet of the dyke. The value of b_0 is calculated in order to conserve the cross-sectional area of the unpressurized dyke, although it can also be specified independently.

In order to decouple the influence of the dyke geometry from the oscillations caused by magma chamber pressure variations, we have assumed a fixed chamber pressure as a boundary condition

for the entrance to the conduit. This assumption is valid because the timescale of chamber pressure variations are much longer than the period of the cycle (2–3 years in comparison with 5–7 weeks).

Results presented in Fig. 8 show that, even with a fixed chamber pressure, there are magma discharge rate oscillations. At the beginning of a cycle, the magma discharge rate is at a minimum, while the overpressure (here presented for 1 km depth by a dashed line) and dyke width are at a maximum. At point A in Fig. 8a, the crystal content and viscosity have reached their maximum values. Beyond this threshold condition, an increase in magma discharge rate results in decreasing pressure and dyke width. However, crystal content and viscosity also decrease, and this effect decreases friction, resulting in flow rate increase and pressure decrease. At C, the system reaches minimum viscosity and crystal content, which cannot decline further. Thereafter the magma discharge rate decreases, while the pressure and dyke width increase. The dyke acts like a capacitor, storing volume during this part of the cycle.

The period of oscillation depends on several parameters such as influx rate and dyke aspect ratio a/R . Typically the period decreases with increasing aspect ratio. The range of calculated periods varies between 38 and 51 days for



Volcanic Eruptions: Cyclicity During Lava Dome Growth, Fig. 8 (a) Dependence of magma discharge rate (solid line) and magmatic overpressure at depth of 1 km (dashed line) on time, for $a = 240$ m and $b = 2.25$ m at the inlet of the dyke. The period of cycle is 46 days; average discharge rate is $6.2 \text{ m}^3/\text{s}$, with peak rate about $12 \text{ m}^3/\text{s}$. (b) Profiles of cross-sectional areas of the conduit during one cycle. Curve A corresponds to the beginning of the cycle, B to a point on the curve of

ascending discharge rate, C to maximum discharge, and D to the middle of descending discharge curve. At the beginning of the cycle, due to large viscosity of magma (at low discharge rate crystal content is high), large magmatic overpressure develops, reaching a maximum near the transition between the dyke and cylindrical conduit; the dyke inflates providing temporary magma storage. Minimum dyke volume corresponds to maximum discharge rate (curve C). (After Costa et al. 2007b)

semimajor axis lengths, a , from 175 to 250 m and semiminor axes, b , from 2 to 4 m. These results match observed cyclicity at SHV. The start of a cycle is quite sharp (Fig. 2), with the onset of shallow hybrid-type (impulsive, low-frequency coda) earthquakes. The change to shorter period and higher amplitude tilt pulsations indicates a marked increase in average magma discharge rate (Voight et al. 1999; Wylie et al. 1999). The model cycles also have rapid onsets. The high amplitude tilt pulsations lasted for several weeks (Voight et al. 1999), consistent with the duration of higher magma discharge rates early in each 5–7-week cycle. Tilt data (Fig. 2) are consistent with the model in that the episode of high magma discharge is associated with a marked deflation that lasts several weeks (see dashed curve at Fig. 8, representing magmatic overpressure at 1 km depth). The magma pressure builds up in the swelling dyke and then reaches a threshold, whereupon a surge of partly crystallized magma occurs, accompanied by elevated seismicity.

The models presented above have certain general features that are necessary to show cyclic behavior. First of all, the resistance of the conduit must depend on magma discharge rate in a way that resistance decreases when discharge rate increases in some range of discharge rate. This dependence is reproduced by a sigmoidal curve. Resistance is a product of viscosity and velocity and is linearly proportional to discharge rate. This means that magma viscosity must decrease as discharge rate strongly increases. There may be many reasons for this behavior, including crystallization, temperature variation, or gas diffusion. The second condition is that there must be some capacitor in the system that can store magma in a period of low discharge rate and release it in a period of high discharge rate. The role of this capacitor can be played by a magma chamber or dyke-shaped conduit located in elastic rocks or by compressibility of the magma itself. The volumes of these capacitors are different and thus will cause pulsations with different periods. Currently

there is no single model that can account for pulsations with multiple timescales.

Future Directions

Our models indicate that magmatic systems in lava dome eruptions can be very sensitive to small changes in parameters. This sensitivity is most marked when the system is close to the cusps of steady-state solutions. If magma discharge rate becomes so high that gas cannot escape efficiently during magma ascent, then conditions for magma fragmentation and explosive eruption can arise. Empirical evidence suggests that conditions for explosive eruption arise when magma discharge rates reach approximately $10 \text{ m}^3/\text{s}$ or more in dome eruptions (Jaupart and Allegre 1991; Sparks 1997). Calculations show the possibility of such high discharge rates for the system parameters typical of lava dome-building eruptions.

We have illustrated model sensitivity of results by varying only one parameter at a time on plots of chamber pressure and discharge rate. However, magmatic systems have many controlling parameters that may vary simultaneously. Furthermore, some controlling parameters are likely to be interdependent (such as temperature, volatile content, and phenocryst content), and others may be independent (such as magma temperature and conduit dimensions). An eruption can be expected to move through n -parameter space, making simulation and its parameter depiction difficult. Our results are simplified, so system sensitivity and behavior in the real world may be yet more complex. A volcanic system may be quite predictable when it is within a stable regime but may become inherently unpredictable (Sparks 2003; Sparks and Aspinall 2004) when variations in the parameters move the system toward transition points and flow regime boundaries.

As in all complex systems, there are many controlling parameters. Our models capture some of the key dynamics but are still simplified in many respects, so do not fully capture the real variations. Our models do not, for example, consider variations in dome height, gas escape to surrounding rocks, strain rate-dependent

rheological effects, or time-dependent changes in conduit diameter. The model for porosity is based on interpretation of measurements of porosity of erupted magma. The role of post-eruptive alterations of pore structure, for example, formation of cooling cracks, cannot be easily estimated. The model of bubble coalescence and permeability formation is important for understanding gas escape mechanisms and will provide constraints on transitions between extrusive and explosive activity. Because the model remains 1D, lateral distribution of parameters cannot be studied. These include lateral pressure gradients, magma crystallization on the conduit walls, wall rock melting or erosion, formation of shear zones and shear heating, and heat flux to surrounding rocks. The models also make the simplifying assumption that influx into the chamber from a deep source is a constant or given as a function of time. The dynamics of the magma chamber itself are oversimplified in all existing conduit flow models. Changes in magma properties in magma chamber can affect the long-term evolution of eruptions. We have considered water as the only volatile, and the addition of other gas species (e.g., CO_2 and SO_2) would add further variability.

There are large uncertainties in some parameters, which are likely to be very strong controls, such as the rheological properties of high-crystalline magmas and crystal growth kinetic parameters, notably at low pressures ($<30 \text{ MPa}$) where experiments are very difficult to do (e.g., Couch et al. 2001). More experiments are necessary to understand the rheology of multiphase systems containing melt, crystals, and bubbles. The effects of crystal shape, crystal size distribution, and strain rate remain largely unclear. Some parameters, such as conduit geometry variation with depth, are highly uncertain. With so many parameters, good fits can be achieved by selecting plausible values for real systems. Barmin et al. (2002), for example, were able to reproduce the patterns of discharge rate at Mount St. Helens and Santiaguito quite accurately. However, such models are not unique, partly because the actual values of some parameters may be quite different to the assumed values and partly because of the model simplifications.

Results obtained from waveform inversions of very long-period seismic data over the past few years point to the predominance of a crack-like geometry for volcanic conduits (Chouet et al. 2005), and it is becoming increasingly evident that the details of this geometry, such as conduit inclination, a sudden change in conduit direction, a conduit bifurcation, or a sudden increase in cross section, are all critically important in controlling magma flow dynamics. Future modeling attempts will need to be closely tied to information on conduit geometry derived from seismology to provide a more realistic view of volcanic processes.

The full simulation of any particular volcanic eruption in such a nonlinear and sensitive system may appear a hopeless task. However, some reduction in uncertainties will certainly help to make the models more realistic. Further experimental studies of crystallization kinetics and the rheological properties of magma at high crystallinities are among the most obvious topics for future research. Advances in understanding the controls on magma input into an open-system chamber would be beneficial, since the delicate balance between input and output is a prime control on periodic behavior.

Further model development includes 2D effects, elastic deformation effects in dyke-fed domes and coupling between magma chamber and conduit flow dynamics. Even with such improvements, large parameter uncertainties and modeling difficulties will remain. In such circumstances the logical approach is to start quantifying the uncertainties and sampling from them to produce probabilistic outputs based on ensemble models where numerical models of the kind discussed here can be run many times. A future challenge for numerical models will also be to produce simulated outputs which compare in detail with observations, in particular time series of magma discharge rates.

Acknowledgments This work was supported by NERC research grant reference NE/C509958/1. OM and AB acknowledge Russian Foundation for Basic Research (08-01-00016) and President of Russian Federation program (NCH-4710.2006.1). RSJS acknowledges a Royal Society Wolfson Merit Award. The Royal Society

exchange grants and NERC grants had supported the Bristol/Moscow work over the last 10 years.

Bibliography

Primary Literature

- Balmforth NJ, Burbidge AS, Craster RV (2001) Shallow lava theory. In: Balmforth NJ, Provenzale A (eds) *Geomorphological fluid mechanics. Lecture notes in physics*, vol 582. Springer, Berlin, pp 164–187
- Balmforth NJ, Burbidge AS, Craster RV, Rust AC, Sassi R (2006) Viscoplastic flow over an inclined surface. *J Non-Newtonian Fluid Mech* 139:103–127
- Barmin A, Melnik O, Sparks RSJ (2002) Periodic behaviour in lava dome eruptions. *Earth Planet Sci Lett* 199:173–184
- Bingham EC (1922) *Fluidity and plasticity*. McGraw-Hill, New York, p 215
- Blake S (1990) Viscoplastic models of lava domes. In: Fink JH (ed) *Lava flows and domes; emplacement mechanisms and hazard implications*, vol 2. Springer, Berlin, pp 88–126
- Blundy JD, Cashman KV, Humphreys MCS (2006) Magma heating by decompression-driven crystallisation beneath andesite volcanoes. *Nature* 443:76–80. <https://doi.org/10.1038/nature05100>
- Calder ES, Luckett R, Sparks RSJ, Voight B (2002) Mechanisms of lava dome instability and generation of rock-falls and pyroclastic flows at Soufrière Hills Volcano, Montserrat. In: Druitt TH, Kokelaar BP (eds) *The eruption of Soufrière Hills Volcano, Montserrat, from 1995 to 1999. Memoirs*, vol 21. Geological Society, London, pp 173–190
- Calder ES, Cortés JA, Palma JL, Luckett R (2005) Probabilistic analysis of rockfall frequencies during an andesite lava dome eruption: the Soufrière Hills Volcano, Montserrat. *Geophys Res Lett* 32:L16309. <https://doi.org/10.1029/2005GL023594>
- Caricchi L, Burlini L, Ulmer P, Gerya T, Vassalli M, Papale P (2007) Non-Newtonian rheology of crystal-bearing magmas and implications for magma ascent dynamics. *Earth Planet Sci Lett* 264:402. <https://doi.org/10.1016/j.epsl.2007.09.032>
- Cashman KV, Blundy JD (2000) Degassing and crystallization of ascending andesite and dacite. In: Francis P, Neubeck J, Sparks RSJ (eds) *Causes and consequences of eruptions of andesite volcanoes. Philosophical transactions of the royal society of London A*, vol 358, pp 1487–1513. The Royal Society, London
- Chouet B, Dawson P, Arciniega-Ceballos A (2005) Source mechanism of Vulcanian degassing at Popocatepetl Volcano, Mexico, determined from waveform inversions of very long period signals. *J Geophys Res* 110: B07301. <https://doi.org/10.1029/2004JB003524>
- Christiansen RL, Peterson DW (1981) Chronology of the 1980 eruptive activity. In: Lipman PW, Mullineaux DR (eds) *The 1980 eruptions of Mount St. Helens*.

- Washington, US geological survey professional paper 1250, p 844. USGS Publications Warehouse. <http://pubs.er.usgs.gov/publication/pp1250>
- Clarke AB, Stephens S, Teasdale R, Sparks RSJ, Diller K (2007) Petrological constraints on the decompression history of magma prior to Vulcanian explosions at the Soufrière Hills volcano, Montserrat. *J Volcanol Geotherm Res* 161:261–274. <https://doi.org/10.1016/j.jvolgeores.2006.11.007>
- Cole P, Calder ES, Sparks RSJ, Clarke AB, Druitt TH, Young SR, Herd R, Harford CL, Norton GE (2002) Deposits from dome-collapse and fountain-collapse pyroclastic flows at Soufrière Hills Volcano, Montserrat. In: Druitt TH, Kokelaar BP (eds) The eruption of the Soufrière Hills Volcano, Montserrat from 1995 to 1999. Memoir, vol 21. Geological Society, London, pp 231–262
- Connor CB, Sparks RSJ, Mason RM, Bonadonna C, Young SR (2003) Exploring links between physical and probabilistic models of volcanic eruptions: the Soufrière Hills Volcano. Montserrat *Geophys Res Lett* 30. <https://doi.org/10.1029/2003GL017384>
- Costa A (2005) Viscosity of high crystal content melts: dependence on solid fraction. *Geophys Res Lett* 32: L22308. <https://doi.org/10.1029/2005GL02430>
- Costa A, Macedonio G (2002) Nonlinear phenomena in fluids with temperature-dependent viscosity: an hysteresis model for magma flow in conduits. *Geophys Res Lett* 29(10):40–1. <https://doi.org/10.1029/2001GL014493>
- Costa A, Macedonio G (2003) Viscous heating in fluids with temperature-dependent viscosity: implications for magma flows. *Nonlinear Proc Geophys* 10:545–555
- Costa A, Macedonio G (2005) Viscous heating in fluids with temperature-dependent viscosity: triggering of secondary flows. *J Fluid Mech* 540:21–38
- Costa A, Melnik O, Sparks RSJ (2007a) Controls of conduit geometry and wallrock elasticity on lava dome eruptions. *Earth Planet Sci Lett* 260:137–151. <https://doi.org/10.1016/j.epsl.2007.05.024>
- Costa A, Melnik O, Sparks RSJ, Voight B (2007b) The control of magma flow in dykes on cyclic lava dome extrusion. *Geophys Res Lett* 34:L02303. <https://doi.org/10.1029/2006GL027466>
- Couch S, Sparks RSJ, Carroll MR (2001) Mineral disequilibrium in lavas explained by convective self-mixing in open magma chambers. *Nature* 411:1037–1039
- Denlinger RP, Hoblitt RP (1999) Cyclic eruptive behaviour of silicic volcanoes. *Geology* 27(5):459–462
- Diller K, Clarke AB, Voight B, Neri A (2006) Mechanisms of conduit plug formation: implications for vulcanian explosions. *Geophys Res Lett* 33:L20302. <https://doi.org/10.1029/2006GL027391>
- Dirksen O, Humphreys MCS, Pletchov P, Melnik O, Demyanchuk Y, Sparks RSJ, Mahony S (2006) The 2001–2004 dome-forming eruption of Shiveluch Volcano, Kamchatka: observation, petrological investigation and numerical modelling. *J Volcanol Geotherm Res* 155:201–226. <https://doi.org/10.1016/j.jvolgeores.2006.03.029>
- Druitt TH, Young S, Baptie B, Calder E, Clarke AB, Cole P, Harford C, Herd R, Luckett R, Ryan G, Sparks RSJ, Voight B (2002) Episodes of cyclic Vulcanian explosive activity with fountain collapse at Soufrière Hills volcano, Montserrat. In: Druitt TH, Kokelaar BP (eds) The eruption of the Soufrière Hills volcano, Montserrat from 1995 to 1999. Memoir, vol 21. Geological Society, London, pp 231–262
- Eichelberger JC, Carrigan CR, Westrich HR, Price RH (1986) Non-explosive silicic volcanism. *Nature* 323:598–602
- Fedotov SA, Dvigalo VN, Zharinov NA, Ivanov VV, Seliverstov NI, Khubunaya SA, Demyanchuk YV, Markov LG, Osipenko LG, Smelov NP (2001) The eruption of Shiveluch volcano on May–July 2001. *Volcanol Seis* 6:3–15
- Fink JH, Griffiths RW (1990) Radial spreading of viscous gravity currents with solidifying crust. *J Fluid Mech* 221:485–509
- Fink JH, Griffiths RW (1998) Morphology, eruption rates, and rheology of lava domes: insights from laboratory models. *J Geophys Res* 103:527–545
- Green DN, Neuberg J (2006) Waveform classification of volcanic low-frequency earthquake swarms and its implication at Soufrière Hills Volcano, Montserrat. *J Volcanol Geotherm Res* 153:51–63. <https://doi.org/10.1016/j.jvolgeores.2005.08.003>
- Hale AJ, Bourguoin L, Mühlhaus HB (2007) Using the level set method to model endogenous lava dome growth. *J Geophys Res* 112:B03213. <https://doi.org/10.1029/2006JB004445>
- Hale AJ, Wadge G (2003) Numerical modeling of the growth dynamics of a simple silicic lava dome. *Geophys Res Lett* 30(19). <https://doi.org/10.1029/2003GL018182>
- Hammer JE, Rutherford MJ (2002) An experimental study of the kinetics of decompression-induced crystallization in silicic melt. *J Geophys Res* 107(B1):ECV 8-1. <https://doi.org/10.1029/2001JB000281>
- Harris AL, Rose WI, Flynn LP (2002) Temporal trends in lava dome extrusion at Santiaguito 1922–2000. *Bull Volcanol* 65:77–89
- Hess KU, Dingwell DB (1996) Viscosities of hydrous leucogranite melts: a non-Arrhenian model. *Am Mineral* 81:1297–1300
- Hoblitt RP, Wolfe EW, Scott WE, Couchman MR, Pallister JS, Javier D (1996) The preclimactic eruptions of Mount Pinatubo, June 1991. In: Newhall CG, Punongbayan RS (eds) Fire and mud: eruptions and lahars of mount Pinatubo, Philippines. Philippine Institute of Volcanology and Seismology/University of Washington Press, Quezon City/Seattle, pp 457–511
- Hort M (1998) Abrupt change in magma liquidus temperature because of volatile loss or magma mixing: effects of nucleation, crystal growth and thermal history of the magma. *J Petrol* 39:1063–1076
- Humphreys M, Blundy JD, Sparks RSJ (2006) Magma evolution and open-system processes at Shiveluch volcano: insights from phenocryst zoning. *J Petrol* 47(12):2303–2334. <https://doi.org/10.1093/petrology/egl045>

- Huppert HE, Shepherd JB, Sigurdsson H, Sparks RSJ (1982) On lava dome growth, with application to the 1979 lava extrusion of the Soufriere, St Vincent. *J Volcanol Geotherm Res* 14:199–222
- Huppert HE, Woods AW (2002) The role of volatiles in magma chamber dynamics. *Nature* 420:493–495
- Ida Y (1996) Cyclic fluid effusion accompanied by pressure change: implication for volcanic eruptions and tremor. *Geophys Res Lett* 23:1457–1460
- Iverson RM et al (2006) Dynamics of seismogenic volcanic extrusion at Mount St. Helens in 2004–2005. *Nature* 444:439–443
- Jaquet O, Sparks RSJ, Carniel R (2006) Magma memory recorded by statistics of volcanic explosions at the Soufriere Hills Volcano, Montserrat. In: Mader HM, Coles SG, Connor CB, Connor LJ (eds) *Statistics in volcanology*. Special publication of IAVCEI, vol 1. Geological Society, London, pp 175–184
- Jaupart C, Allegre CJ (1991) Gas content, eruption rate and instabilities of eruption regime in silicic volcanoes. *Earth Planet Sci Lett* 102:413–429
- Kirkpatrick R (1976) Towards a kinetic model for the crystallization of magma bodies. *J Geophys Res* 81:2565–2571
- Landau L, Lifshitz E (1987) *Fluid mechanics*, 2nd edn. Butterworth-Heinemann, Oxford
- Lejeune A, Richet P (1995) Rheology of crystal-bearing silicate melts: an experimental study at high viscosity. *J Geophys Res* 100:4215–4229
- Lensky NG, Sparks RSJ, Navon O, Lyakhovsky V (2007) Cyclic activity at Soufriere Hills volcano, Montserrat: degassing-induced pressurization and stick-slip extrusion. In: Lane SJ, Gilbert JS (eds) *Fluid motions in volcanic conduits: a source of seismic and acoustic signals*. Special publications. The geological society of London 2008, vol 307. Geological Society, London, pp 169–188. [https://doi.org/10.1144/SP307.100305-8719/08/\\$15.00](https://doi.org/10.1144/SP307.100305-8719/08/$15.00)
- Lister JR, Kerr RC (1991) Fluid mechanical models of crack propagation and their application to magma transport in dykes. *J Geophys Res* 96:10049–10077
- Llewellyn EW, Manga M (2005) Bubble suspension rheology and implications for conduit flow. *J Geotherm Res* 143:205–217
- Loitsyansky LG (1978) *Fluid and gas mechanics*. Nauka, Moscow, p 847. (in Russian)
- Maeda I (2000) Nonlinear visco-elastic volcanic model and its application to the recent eruption of Mt. Unzen. *J Volcanol Geotherm Res* 95:35–47
- Mason RM, Starostin AB, Melnik O, Sparks RSJ (2006) From Vulcanian explosions to sustained explosive eruptions: the role of diffusive mass transfer in conduit flow dynamics. *J Volcanol Geotherm Res* 153:148–165. <https://doi.org/10.1016/j.jvolgeores.2005.08.011>
- Marsh BD (2000) Reservoirs of magma and magma chambers. In: Sigurdsson H (ed) *Encyclopedia of volcanoes*. Academic Press, New York, pp 191–206
- Mastin GL, Pollard DD (1988) Surface deformation and shallow dike intrusion processes at Inyo craters, Long Valley, California. *J Geophys Res* 93(B11):13221–13235
- Matthews SJ, Gardeweg MC, Sparks RSJ (1997) The 1984 to 1996 cyclic activity of Lascar Volcano, northern Chile: cycles of dome growth, dome subsidence, degassing and explosive eruptions. *Bull Volcanol* 59:72–82
- Mattioli G, Dixon TH, Farina F, Howell ES, Jansma PE, Smith AL (1998) GPS measurement of surface deformation around Soufriere Hills volcano, Montserrat from October 1995 to July 1996. *Geophys Res Lett* 25(18):3417–3420
- Melnik O (2000) Dynamics of two-phase conduit flow of high-viscosity gas-saturated magma: large variations of sustained explosive eruption intensity. *Bull Volcanol* 62:153–170
- Melnik O, Barmin A, Sparks RSJ (2005) Dynamics of magma flow inside volcanic conduits with bubble overpressure buildup and gas loss through permeable magma. *J Volcanol Geotherm Res* 143:53–68
- Melnik O, Sparks RSJ (1999) Non-linear dynamics of lava dome extrusion. *Nature* 402:37–41
- Melnik O, Sparks RSJ (2002) Dynamics of magma ascent and lava extrusion at Soufriere Hills Volcano, Montserrat. In: Druitt TH, Kokelaar BP (eds) *The eruption of the Soufriere Hills Volcano, Montserrat from 1995 to 1999*. Memoir, vol 21. Geological Society, London, pp 223–240
- Melnik O, Sparks RSJ (2005) Controls on conduit magma flow dynamics during lava dome building eruptions. *J Geophys Res* 110(B02209). <https://doi.org/10.1029/2004JB003183>
- Mériaux C, Jaupart C (1995) Simple fluid dynamic models of volcanic rift zones. *Earth Planet Sci Lett* 136:223–240
- Murphy MD, Sparks SJ, Barclay J, Carroll MR, Brewer TS (2000) Remobilization origin for andesite magma by intrusion of mafic magma at the Soufriere Hills Volcano. *J Petrol* 41:21–42. Montserrat WI (ed) *A trigger for renewed eruption*
- Muskhelishvili N (1963) *Some basic problems in the mathematical theory of elasticity*. Noordhoff, Leiden
- Nakada S, Eichelberger JC (2004) Looking into a volcano: drilling Unzen. *Geotimes* 49:14–17
- Nakada S, Shimizu H, Ohta K (1999) Overview of the 1990–1995 eruption at Unzen Volcano. *J Volcanol Geoth Res* 89:1–22
- Navon O, Lyakhovsky V (1998) Vesiculation processes in silicic magmas. In: Gilbert J, Sparks RSJ (eds) *The physics of explosive volcanic eruption*. Special publication, vol 145. Geological Society, London, pp 27–50
- Neuberg JW, Tuffen H, Collier L, Green D, Powell T, Dingwell D (2006) The trigger mechanism of low-frequency earthquakes on Montserrat. *J Volcanol Geotherm Res* 153:37–50

- Newhall CG, Melson WG (1983) Explosive activity associated with the growth of volcanic domes. *J Volcanol Geotherm Res* 17:111–131
- Norton GE, Watts RB, Voight B, Mattioli GS, Herd RA, Young SR, Devine JD, Aspinall WP, Bonadonna C, Baptie BJ, Edmonds M, Harford CL, Jolly AD, Loughlin SC, Luckett R, Sparks RSJ (2002) Pyroclastic flow and explosive activity of the lava dome of Soufrière Hills volcano, Montserrat, during a period of no magma extrusion (March 1998 to November 1999). In: Druitt TH, Kokelaar BP (eds) *The eruption of the Soufrière Hills volcano, Montserrat from 1995 to 1999. Memoir, vol 21. Geological Society, London*, pp 467–482
- Ohba T, Kitade Y (2005) Subvolcanic hydrothermal systems: implications from hydrothermal minerals in hydrovolcanic ash. *J Volcanol Geotherm Res* 145:249–262
- Robertson R, Cole P, Sparks RSJ, Harford C, Lejeune AM, McGuire WJ, Miller AD, Murphy MD, Norton G, Stevens NF, Young SR (1998) The explosive eruption of Soufrière Hills Volcano, Montserrat 17 September, 1996. *Geophys Res Lett* 25:3429–3432
- Roman DC (2005) Numerical models of volcanotectonic earthquake triggering on non-ideally oriented faults. *Geophys Res Lett* 32. <https://doi.org/10.1029/2004GL021549>
- Roman DC, Neuberg J, Luckett RR (2006) Assessing the likelihood of volcanic eruption through analysis of volcanotectonic earthquake fault-plane solutions. *Earth Planet Sci Lett* 248:244–252
- Rubin AM (1995) Propagation of magma-filled cracks. *Annu Rev Planet Sci* 23:287–336
- Saar MO, Manga M, Katharine VC, Fremouw S (2001) Numerical models of the onset of yield strength in crystal-melt suspensions. *Earth Planet Sci Lett* 187:367–379
- Sahagian D (2005) Volcanic eruption mechanisms: insights from intercomparison of models of conduit processes. *J Volcanol Geotherm Res* 143(1–3):1–15
- Slezin YB (1984) Dispersion regime dynamics in volcanic eruptions, 2. Flow rate instability conditions and nature of catastrophic explosive eruptions. *Vulkanol Seism* 1:23–35
- Slezin YB (2003) The mechanism of volcanic eruptions (a steady state approach). *J Volcanol Geotherm Res* 122:7–50
- Sparks RSJ (1978) The dynamics of bubble formation and growth in magmas – a review and analysis. *J Volcanol Geotherm Res* 3:1–37
- Sparks RSJ (1997) Causes and consequences of pressurization in lava dome eruptions. *Earth Planet Sci Lett* 150:177–189
- Sparks RSJ (2003) Forecasting volcanic eruptions. *Earth Planet Sci Lett Front Earth Sci Ser* 210:1–15
- Sparks RSJ, Aspinall WP (2004) Volcanic activity: frontiers and challenges. In: *Forecasting, prediction, and risk assessment. AGU geophysical monograph “state of the planet”. IUGG monograph 19, vol 150, pp 359–374. American Geophysical Union, Washington DC*
- Sparks RSJ, Murphy MD, Lejeune AM, Watts RB, Barclay J, Young SR (2000) Control on the emplacement of the andesite lava dome of the Soufrière Hills Volcano by degassing-induced crystallization. *Terra Nova* 12:14–20
- Sparks RSJ, Young SR (2002) The eruption of Soufrière Hills volcano, Montserrat (1995–1999): overview of scientific results. In: Druitt TH, Kokelaar BP (eds) *The eruption of the Soufrière Hills Volcano, Montserrat from 1995 to 1999. Memoir, vol 21. Geological Society, London*, pp 45–69
- Sparks RSJ, Young SR, Barclay J, Calder ES, Cole PD, Darroux B, Davies MA, Druitt TH, Harford CL, Herd R, James M, Lejeune AM, Loughlin S, Norton G, Skeritt G, Stevens NF, Toothill J, Wadge G, Watts R (1998) Magma production and growth of the lava dome of the Soufrière Hills Volcano, Montserrat, West Indies: November 1995 to December 1997. *Geophys Res Lett* 25:3421–3424
- Swanson DA, Holcomb RT (1990) Regularities in growth of the Mount St. Helens dacite dome 1980–1986. In: Fink JH (ed) *Lava flows and domes; emplacement mechanisms and hazard implications. Springer, Berlin*, pp 3–24
- Voight B, Hoblitt RP, Clarke AB, Lockhart AB, Miller AD, Lynch L, McMahon J (1998) Remarkable cyclic ground deformation monitored in real-time on Montserrat, and its use in eruption forecasting. *Geophys Res Lett* 25:3405–3408
- Voight B, Sparks RSJ, Miller AD, Stewart RC, Hoblitt RP, Clarke A, Ewart J, Aspinall W, Baptie B, Druitt TH, Herd R, Jackson P, Lockhart AB, Loughlin SC, Lynch L, McMahon J, Norton GE, Robertson R, Watson IM, Young SR (1999) Magma flow instability and cyclic activity at Soufrière Hills Volcano, Montserrat. *Science* 283:1138–1142
- Walker GPL (1973) Lengths of lava flows. *Philos Trans R Soc A* 274:107–118
- Watson IM et al (2000) The relationship between degassing and ground deformation at Soufrière Hills Volcano, Montserrat. *J Volcanol Geotherm Res* 98(1–4):117–126
- Watts RB, Sparks RSJ, Herd RA, Young SR (2002) Growth patterns and emplacement of the andesitic lava dome at Soufrière Hills Volcano, Montserrat. In: Druitt TH, Kokelaar BP (eds) *The eruption of the Soufrière Hills Volcano, Montserrat from 1995 to 1999. Memoir, vol 21. Geological Society, London*, pp 115–152
- Whitehead JA, Helfrich KR (1991) Instability of flow with temperature-dependent viscosity: a model of magma dynamics. *J Geophys Res* 96:4145–4155
- Williams SN, Self S (1983) The October 1902 Plinian eruption of Santa Maria volcano, Guatemala. *J Volcanol Geotherm Res* 16:33–56

- Woods AW, Koyaguchi T (1994) Transitions between explosive and effusive eruption of silicic magmas. *Nature* 370:641–645
- Wylie JJ, Voight B, Whitehead JA (1999) Instability of magma flow from volatile-dependent viscosity. *Science* 285:1883–1885
- Yokoyama I, Yamashita H, Watanabe H, Okada H (1981) Geophysical characteristics of dacite volcanism – 1977–1978 eruption of Usu volcano. *J Volcanol Geotherm Res* 9:335–358

Books and Reviews

- Dobran F (2001) *Volcanic processes: mechanisms in material transport*. Kluwer, New York, p 620
- Gilbert JS, Sparks RSJ (eds) (1998) *The physics of explosive volcanism*. *Spec Publ Geol Soc Lond* 145:186
- Gonnermann H, Manga M (2007) The fluid mechanics inside a volcano. *Ann Rev Fluid Mech* 39:321–356
- Mader HM, Coles SG, Connor CB, Connor LJ (2006) *Statistics in volcanology*. IAVCEI publications. Geological Society Publishing House, p 296. Geological Society, London

Volcanic Eruptions: Stochastic Models of Occurrence Patterns

Mark S. Bebbington
Massey University, Palmerston North,
New Zealand

Article Outline

Glossary
Definition of the Subject
Introduction
Data
Temporal Models
Volcanic Regimes
Spatial Aspects
Yucca Mountain
Interactions with Earthquakes
Model Assessment
Future Directions
Bibliography

Keywords

Temporal Models · Spatial Models · Fractals ·
Volcanic Regimes · Interactions with
Earthquakes · Model Assessment and Fit

Glossary

Absolute time is denoted by t or s , with the latter usually being the time of the last known event.

The **elapsed time** since the last known event is denoted by u , or, equivalently, $t - s$.

A **parameter estimate** of the parameter λ , say, is denoted by $\hat{\lambda}$.

Inter-onset time Time between successive onsets, denoted by $r_i = t_{i+1} - t_i$, for $i = 1, \dots, n-1$. In many papers this is termed the **repose time**, which we shall use interchangeably, although

the latter is strictly the time between the *end* of one eruption and the *start* of the next.

Onset The beginning of an eruption. The term **event** will be used interchangeably.

Onset time Time at which an eruption begins. The i th onset time will be denoted t_i , for $i = 1, \dots, n$.

Polygenetic vent Site of multiple events, in contrast to **monogenetic vents**, at which only one eruption occurs. The latter occur predominately in **volcanic fields**.

Repose Periods during which an eruption is not in progress.

Definition of the Subject

Depending on the definition, there are approximately 1,300 Holocene (last 10,000 years) active volcanoes (Simkin 1993), some 55–70 of which are typically active in any given year. Being tectonic phenomena, they are spatially overrepresented along littorals and in island arcs where populations congregate, and millions of lives are potentially at risk. Hence, the quantitative forecasting of hazard has long been a key aim of volcanology. This requires models to quantify the likelihood of future outcomes, which may involve the size, location, and/or style of eruption, in addition to temporal occurrence. While a wide range of complex deterministic models exist to model various volcanic processes, these provide little in the way of information about future activity. Being the (partially) observed realization of a complex system, volcanological data are inherently stochastic in nature and need to be modeled using statistical models. To paraphrase Decker (1986) concerning potentially active volcanoes, geologists can find out what did happen, and geophysicists can determine what is happening, but to find out what might happen requires statistics. Reminding ourselves that “all models are wrong, but some are useful” (Box and Draper 1987), the most useful models for volcanic eruptions are ones that provide new insight into the physical

processes and can in turn be informed by our present understanding of those physical processes.

Introduction

The hazard is a product of many elements, including the likelihood of an eruption, the distribution of the eruption size, and the style of the eruption. Depending on the nature of the volcanic products, climatological or geographical factors may also play a part.

The style of an eruption strongly determines the relative risks to life and property. Effusive eruptions of molten lava pose little threat to life with, for example, only one life being lost due to a volcanic eruption on Hawaii in the twentieth century. However, in the same time period, 5 % of the island has been covered by new lava flows (Decker 1986). Explosive eruptions, in particular pyroclastic flows, travel at high speeds and can cause extensive fatalities such as the 29,000 killed in the Mount Pelée (West Indies) eruption of 1902. Lahars, or mudflows, such as those that killed 25,000 people in the Nevado del Ruiz (Colombia) eruption of 1985, are also a major danger.

Forecasting volcanic eruptions is easier than forecasting earthquakes on several grounds. First, at least in the case of polygenetic volcanoes, there need be no spatial element involved. Second, it is possible using geological techniques to look backward (often a long way, depending on geography and finances) to see what happened in the past. Thirdly, and of most importance for short-term warning, magma intrusions can be detected by seismic networks, although not all seismic signals indicate magma of course, and other monitoring techniques (Decker 1986).

In this entry we will largely limit our consideration to models of eruption occurrence, omitting techniques for forecasting the nature and effect of the eruption. For an introduction to the latter, see Orsi et al. (2011). As the track record of a potentially active volcano provides the best method of assessing its future volcanic hazards on a long-term basis (Crandell et al. 1975; Decker 1986), we

will first briefly review the provenance and characteristics of the data available. Various taxonomies for stochastic models are possible. As the majority of models applied to volcanoes are purely temporal, and this also includes the simplest models, we will look at these first. Such models can be further broken down into (nonhomogeneous) Poisson processes, which consider absolute time relative to some fixed origin, and renewal processes, which consider only the time since the previous event. Markov processes, where the state of the process moves between various eruptive and repose states, and models based on eruptive volume are the simplest ways of incorporating some element of eruption size into the model. Fractals have been used as a descriptive technique for volcanoes and are reviewed next. The above models, with the exception of the nonhomogeneous Poisson process, all assume that the volcano is in steady state, i.e., that the activity pattern does not change over time. For volcanoes with lengthy records, this appears to be an untenable assumption over longer timescales (Wadge 1982). This raises the issues of, first, how to detect a departure from stationary stochastic behavior or change in volcanic regime and, secondly, how to model it, which are discussed in section “[Volcanic Regimes](#)” illustrated with results from Mount Etna. While purely temporal models are suitable for polygenetic volcanoes, monogenetic volcanic fields require a spatial element, to model the location, as well as the time, of future events. This is usually estimated using kernel smoothing, but in cases where the location can be treated as a discrete variable, such as summit versus flank eruptions, Markov chain techniques can be used. These are reviewed, along with methods for detecting clustering and linear alignments, in section “[Spatial Aspects](#).” A very high profile, particularly in the USA, exercise in volcanic hazard estimation concerned the risk to a proposed high-level radioactive waste repository planned for Yucca Mountain. The history of published hazard estimates is used in section “[Yucca Mountain](#)” to illustrate the application of both temporal and spatiotemporal models. Section “[Interactions with Earthquakes](#)” looks at methods for detecting links between volcanic

eruptions and large earthquakes. Determining whether correlation in the occurrence patterns exists is a necessary precursor to formulating models to take advantage of any potential new information from the additional process. The final substantive section looks at the question of model assessment, or how models can be validated.

The focus of this entry is squarely on the techniques. Hence, the applications jump around a bit, as they are those presented in the original paper (s) dealing with the technique. The examples of Mount Etna and Yucca Mountain were selected for more detailed examination partly on the basis that many, somewhat contradictory, results exist. Different models make different assumptions and vary in how much information they can extract from data. In addition, the data used often varies from study to study, and the sensitivity of models to data is important, but too often ignored. Subjective judgments on the merits of various techniques are minimized, except for comments on details necessary to their application.

Data

Information about past eruptions forms the basis of any stochastic model for future behavior. Using such data poses a number of challenges because of its inhomogeneity, particularly in the longer records. This inhomogeneity is due to incomplete observation, which becomes more incomplete the further back one goes, and by variations in dating and measurement precision, again becoming less precise the further into the past one looks. We will now outline the nature and limitations of the available information.

There are several functional distinctions in the types of data used in volcanic hazard models. The most fundamental is at the question of scale. Certain volcanoes, notably Stromboli (which lends its name to the behavior) and Soufriere Hills, erupt in distinct events separated by minutes or hours. Hence, a catalog of hundreds of events may only cover a few days of activity. More commonly, eruptions are events with durations ranging from days to years. Eruptions are usually modeled

simply as a point process of onsets, but a few studies consider duration, volume, or spatial extent.

A global catalog of Holocene volcanism is maintained by the Smithsonian (Siebert and Simkin 2002). This gives onset dates, of various precision, and Volcanic Explosivity Index (VEI) for all known eruptions. The VEI, as a logarithmic measuring scale of eruption size analogous to the magnitude of an earthquake, was proposed by Newhall and Self (1982). It is assigned to historical eruptions on the basis of (in decreasing order of reliability) explosion size, volume of ejecta, column height, classification (“Hawaiian,” “Strombolian,” “Vulcanian,” “(Ultra-)Plinian”), duration, most explosive activity type, and tropospheric and stratospheric injection. Many historical (i.e., observed and recorded at the time) observations also have a duration, and some an estimate of eruptive volume. Estimates of volume, where they exist, typically contain uncertainties of 10–50 % (Wadge and Guest 1981) or more (Martin and Rose 1981), and, provided the eruption styles are similar, the duration of an eruption can be used as a proxy (Bebbington 2007; Mulargia et al. 1985). Other global catalogs include the large quaternary eruption database (Croswell et al. 2012), the completeness and consequent hazard rate of which has been analyzed (Deligne et al. 2010; Furlan 2010), and the caldera collapse database (Geyer and Marti 2008), preliminary analysis of which can be found in Sobradelo et al. (2010).

The frequency of observed volcanic eruptions has increased over the last few hundred years, although this is largely an artifact of the global population increase (Simkin 1994), which indicates that many of the historical records are incomplete. A number of published catalogs for individual volcanoes are of greater detail and probably more complete. Some of the better known are Etna (Behncke et al. 2005; Branca and Del Carlo 2004; Mulargia et al. 1985, 1987; Sandri et al. 2005; Tanguy et al. 2007) and Vesuvius (Carta et al. 1981; Scandone et al. 1993a, 2008) (Italy), Ruapehu (Bebbington and Lai 1996b) and Taranaki (Turner et al. 2011) (New Zealand), Colima (Mexico) (De la Cruz-

Reyna 1993; Medina Martinez 1983; Mendoza-Rosas and De La Cruz-Reyna 2008), and Kilauea/ Mauna Loa (Hawaii) (Klein 1982).

While it is reasonably easy, although sometimes controversial, to define events in an historical catalog, it is not so easy in prehistorical cases. Ideally events correspond to eruptions, but subsequent geological processes can obscure, or even obliterate, the evidence of previous eruptions. The identification of events is dependent on looking in the right place. Prehistorical eruptions can be identified from a combination of stratigraphy (which provides an ordering for the eruptions) and dating techniques. Radiocarbon dating is feasible for material within the last 50,000 years and provides an estimated age and (Gaussian) error given in ^{14}C years. See De la Cruz-Reyna and Carrasco-Nunez (2002) for an example of such a catalog. Because of the variable rate of ^{14}C deposition, this has to be converted to calendar years (Bronk Ramsey 1995), in the process losing the Gaussian distribution (and possibly even becoming disjoint). One way (Turner et al. 2008a) of dealing with the resulting untidiness is to treat the data in Monte Carlo fashion, repeatedly drawing random realizations of the eruption sequence from the distributions (ensuring that the stratigraphic ordering is not violated), and fitting the model to each realization to provide a population of hazard models. Beyond 50,000 years, dating is via a number of other techniques, such as K-Ar radiometric age determinations which, with resolutions of the order of 10^6 – 10^7 years, are much less precise. Conway et al. (1998) take the approach of simulating event dates from a uniform distribution, bounded by the $\pm 2\sigma$ error limits on the dates, in order to determine the uncertainty in the estimated recurrence rate.

Because of the expense, and due to the difficulty of finding suitable material in the sample to date, it is common to date only a portion of the samples. Stratigraphy is then used to interpolate or bound the unknown dates. For example, in a single core, a spline can be fitted to the depth and known dates, representing the sediment deposition rate, and the unknown dates calculated from their depths (Turner et al. 2008a), as shown in Fig. 1. This gets much more complicated when there is a spatial element involved (Condit and

Connor 1996). While deposition cores can provide a complete (above a certain minimum size) and ordered tephra record for a given site, the record extracted from another site may only partially overlap (Burt et al. 2001). Various criteria of tephra composition and geochemistry can be used to identify common events (Turner et al. 2009). Composite records from many sources will almost certainly be incomplete due to missing events (Wang and Bebbington 2012) and may well unknowingly contain duplicates of other events. A method for combining geological and historical catalogs has been proposed by Mendoza-Rosas and De la Cruz-Reyna (2008).

Temporal Models

Models for the occurrence of volcanic eruptions are based on the idea that the past behavior of a volcano is the best predictor of its future activity. This was dramatically illustrated by Crandell et al. (1975) in their semiquantitative forecast of an eruption of Mount St Helens:

The repetitive nature of the eruptive activity at Mount St. Helens during the last 4000 years, with dormant intervals typically of a few centuries or less, suggests that the current quiet period will not last a thousand years. Instead, an eruption is likely within the next hundred years, possibly before the end of this century.

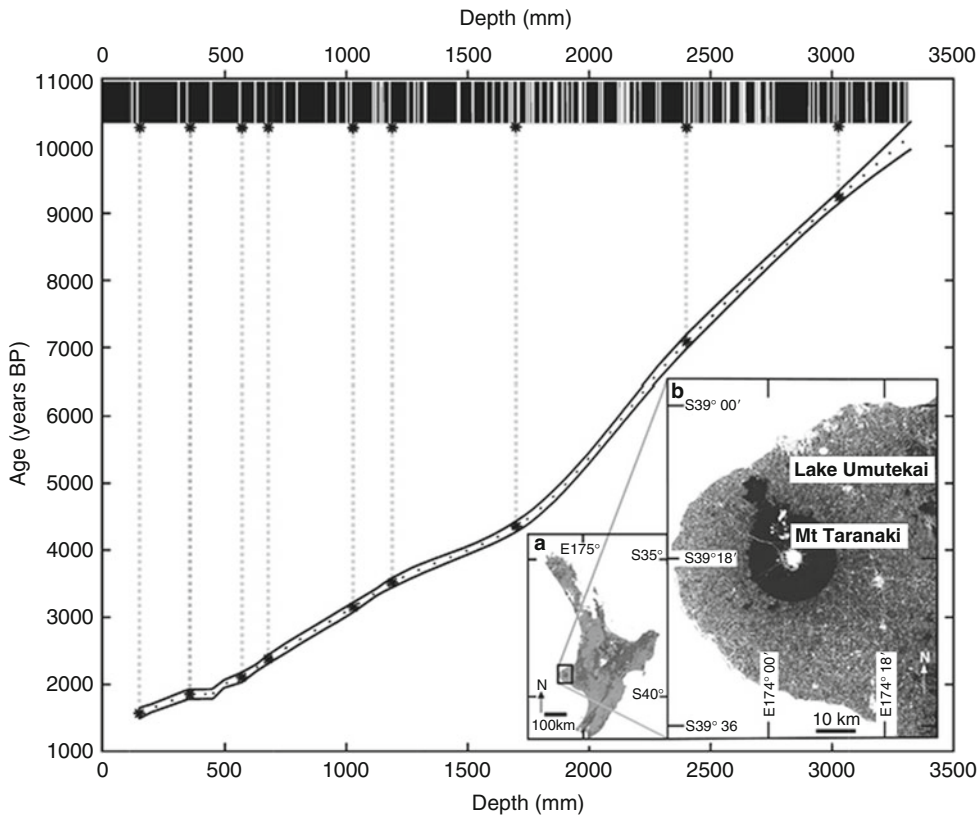
Published in 1975, as a warning that dormant intervals can be a recurring element in the life of a volcano, this foreshadowed the 1980 eruption and subsequent activity.

Let the number of events in a time interval (s, t) be denoted $N(s, t)$, where we use the shorthand $N(t)$ if $s = 0$. The *intensity* of a process is

$$\lambda(t) = \lim_{\Delta t \downarrow 0} \frac{\Pr(N(t, t + \Delta t) = 1)}{\Delta t}, \quad (1)$$

i.e., in the short period of time $(t, t + \Delta t)$, the probability of an event is approximately $\lambda(t)\Delta t$.

Guttorp and Thompson (1991) provided a non-parametric method of estimating the intensity, based on the observed onset counts Y_k in intervals of length Δ , indexed by $k = 1, \dots, K$. They first estimate the reporting probability $p(t)$ by



Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 1 Plot of the radiocarbon age versus depth model (mm) for tephra layers in the sediment core from Lake Umutekai (inset location maps of **a** North Island of New Zealand and **b** Taranaki). The core stratigraphy is

displayed at the top of the plot with *white* layers representing tephra deposits. The locations of radiocarbon dated layers are indicated by stars (Bull Volcanol 70:508, Fig. 1, © Springer-Verlag 2007)

smoothing the observed onset counts (see Solow (1993) for an alternative approach assuming a nondecreasing reporting probability) and calculate $Z_k = Y_k / \int_{(k-1)\Delta}^{k\Delta} \Delta p(t) dt$. This is a reconstruction of the underlying process obtained by reinflating the observed counts by the corresponding observance probability. Assuming that the intensity is locally fairly smooth, an estimate, based on the hanning operation from time-series analysis, is

$$\hat{\lambda}_\lambda(j\Delta) = \frac{\sum_{i=1}^{K-j-1} Z_i Z_{i+j-1} + 2 \sum_{i=1}^{K-j} Z_i Z_{i+j} + \sum_{i=1}^{K-j+1} Z_i Z_{i+j-1}}{4\Delta(K-j) \sum_{k=1}^K Z_k / K} \quad (2)$$

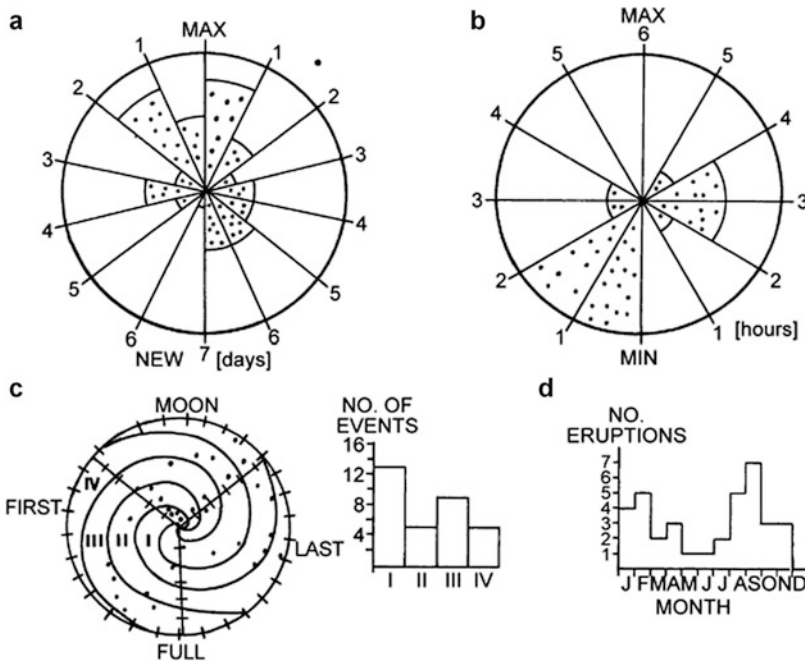
Application to the global catalog, assuming a constant rate of underlying events, produced an estimated observance probability that increases slowly from 1500 until the second half of the

eighteenth century and then more sharply at a constant rate to an assumed probability $p(1980) = 1$. The resulting estimated intensity for global eruption starts did not differ from a Poisson (constant intensity) process. However, in order to see if more structure could be observed in a series of eruptions from a restricted area with homogeneous tectonic structure, the methodology was applied to a catalog of Icelandic eruptions. As expected, the estimated observance probability for the last 200 years was approximately constant, which resulted in a periodicity of about 40 years in the estimated intensity. Alternative methods based on extreme value theory (Deligne et al. 2010; Furlan 2010), which assumes that larger and/or more recent events are more likely to be observed, and hidden Markov models (Wang and Bebbington 2012), where the hidden state is the

number of missed events between observed events, can be applied to global catalogs and individual volcano records, respectively.

There have been a few other attempts to discern periodic fluctuations in volcanic activity. Mauk and Johnston (1973) examined the correlation of the onset dates of 680 eruptions 1900–1971 with the fortnightly solid earth tide. There was a statistically significant tendency for both andesitic and basaltic eruptions to occur at the maxima of the earth tide. In addition, basaltic eruptions had a significant peak at the minima of the earth tide. A more detailed examination of 18 Japanese volcanoes indicated that this may be related to Bouguer anomalies. Similarly, Martin and Rose (1981), in Fig. 2, found that eruption onsets of Fuego volcano, Guatemala, were correlated with the fortnightly lunar tidal maxima, with

23 out of 48 eruptions occurring within ± 2 days of the peak gravity acceleration. Casetti et al. (1981) binned eruptions at Mt Etna and examined the resulting statistics, in particular the month of an eruption. Eruptions were shown to be significantly more likely during March–April and November, when there is an increase in the earth-velocity rate. Although Stothers (1989) repeated this analysis for a global catalog of eruptions with $VEI \geq 3$, 1500–1980, and detected no significant monthly or seasonal variation, Mason et al. (2004) found that seasonal fluctuations account for 18 % of the historical average monthly eruption rate. The idea of a deterministic cycle is incorporated in a renewal model by Jupp et al. (2004), but the result was not fitted to data. This was addressed (Bebbington 2010) in the framework of the trend renewal process, where time is



Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 2 (a) Lambert equal-area polar histogram showing the percentage of eruptions of Fuego since 1800 in relationship to the fortnightly tidal maximum of the vertical gravity acceleration. (b) Lambert equal-area polar histogram showing the percentage of all the known times of the beginning of eruptions since 1957 compared to the 12.4 h (semidiurnal) minimum of the vertical gravity acceleration. (c) A spiral diagram indicating the position of an eruption in relation to the synodic month (days plotted around the circumference) and the reduced anomalistic

age (days plotted as the radius, with perigee at the center), in which it occurred. There are 27.5 days in the anomalistic cycle and 29.6 days in the synodic cycle. Right of the spiral diagram is a histogram of the four spirals of the above diagram, which are drawn around ± 3.5 days of the maxima and minima of the two cycles. (d) Monthly histogram of Fuego's eruptions which have occurred since 1800 (Reprinted from Martin DP, Rose WI (1981) Behavioural patterns of Fuego volcano, Guatemala. J Volcanol Geotherm Res 10:72 © 1981 Elsevier B.V., with permission from Elsevier)

scaled between eruptions, potentially in a periodic fashion. Among other things, the cycles of Fuego (Martin and Rose 1981) and Colima (De la Cruz-Reyna 1993; Luhr and Carmichael 1990) were statistically identified (Fig. 3, where $\lambda(t) \approx \psi(t) N(T)/T$ for observations spanning a time interval of length T).

Poisson Processes

The simplest model is a *simple (homogeneous) Poisson process*, which is characterized by $\lambda(t) = \lambda_0$, a constant independent of t . In this case $N(t)$ has a Poisson distribution with mean $\mu = \lambda_0 t$, and the inter-event times have an exponential distribution with mean $1/\lambda_0$. This is sometimes referred to as *Poisson behavior*. The parameter is estimated as $\hat{\lambda}_0 = N(T)/T$, for an observed history spanning the time interval $(0, T)$.

In order to weaken the requirement that the mean and variance of $N(t)$ be equal, Ho (1990) suggested a Bayesian approach, in which the parameter λ_0 has a gamma prior distribution. This leads to $N(t)$ having a negative binomial distribution. Solow (2001) augmented this to an empirical Bayesian formulation by suggesting that an informative prior could be constructed from the eruption records of a group of similar volcanoes. The negative binomial is of course over-dispersed, and thus, eruption onsets under

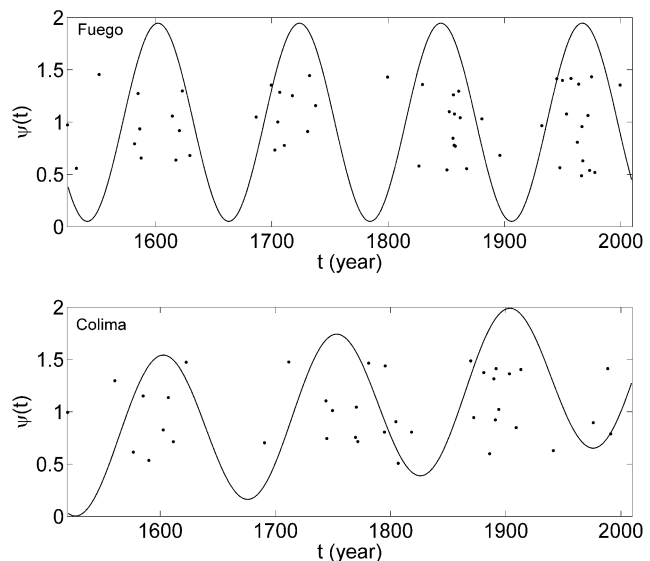
this model are more clustered than in the Poisson process; this constraint was addressed in Rodado et al. (2011), allowing the analog set to be arbitrarily similar to the target volcano. Bebbington and Lai (1998) introduced a generalized negative binomial distribution incorporating serial dependence and applied it to eruptions from Mount Sangay (Ecuador). The Poisson process embedded in the Bayesian framework is also the basis of the Bayesian event tree models for long-term volcanic hazard (Marzocchi et al. 2010; Sobradelo and Marti 2010), although a recent development (Garcia-Aristizabal et al. 2013) allows the incorporation of temporal models with more structure.

The Poisson process is stationary in time, in that the distribution of the number of events in an interval depends only on the length of the interval, not its location. In other words, events occur at the same average rate at all times, i.e., there is no trend in the occurrence rate with time. Marzocchi (1996) analyzed possible trends by calculating the (biased) autocorrelation function

$$a(s) = \frac{\sum_{i=1}^{n-1-|s|} (r_i - \bar{r})(r_{i+|s|} - \bar{r})}{\sum_{i=1}^{n-1} (r_i - \bar{r})^2}, \quad (3)$$

for $s = 0, \pm 1, \dots, \pm m$, where r_i , $i = 1, \dots, n-1$, is the time between the i th and $(i+1)$ th onsets,

Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 3 Fitted cyclic plus trend TRP $\psi(t)$ for (top) Fuego and (bottom) Colima. Data points indicated by dots are the onset dates, jittered vertically (From Bebbington MS (2010) Trends and clustering in the onsets of volcanic eruptions. J Geophys Res 115:B01203 Fig. 6, © 2010 by the American Geophysical Union)



$\bar{r} = \left(1/(n-1) \sum_{i=1}^{n-1} r_i\right)$, and $m < n-1$. Under the null hypothesis that the process is purely random, the distribution of $a(s)$ is asymptotically normal with mean zero and variance $1/(n-1)$ for $s \neq 0$.

More generally, a process can be a *non-homogeneous Poisson process*, where $N(t)$ has a Poisson distribution with mean $\mu(t) = \int_0^t \lambda(s) ds$. Ho (1991) used an example of a non-homogeneous Poisson process known as the *Weibull process* (or *power-law process*) (Bain 1978), with

$$\lambda(t) = \frac{\beta}{\theta} \left(\frac{t}{\theta}\right)^{\beta-1}, \quad (4)$$

where t is absolute time with respect to some fixed origin, not the elapsed time since the previous eruption. This includes the homogeneous Poisson process as a special ($\beta = 1$) case, while if $\beta \neq 1$, the process is nonstationary. The intensity (Eq. 4) is monotonic and hence can model either an increase or decrease in volcanic activity, but not both. Note that the parameter estimates

$$\hat{\beta} = \frac{N(T)}{\sum_{i=1}^{N(T)} \ln(T/t_i)} \quad \text{and} \quad \hat{\theta} = T/N(T)^{1/\beta}, \quad (5)$$

for the intensity (Eq. 4) are sensitive to the position of the time origin (Bebbington and Lai 1996a). Furthermore, estimates of $\hat{\beta} > 2$, which are quite feasible, indicate a constantly accelerating, or convex, intensity. Neither of these properties is particularly desirable from a physical viewpoint, and hence the Weibull process is not suitable to model entire volcanic histories. Salvi et al. (2006) (see also Smethurst et al. 2009) showed statistically that the most recent part of the Mount Etna sequence was nonstationary and fitted the Weibull process to it. The Weibull process can be tested for goodness of fit to a series of data (Bebbington and Lai 1996a) either via a

standard χ^2 test or by using the fact that $x_1, \dots, x_{n-1}$, where $x_i = \ln(t_n/t_{n-i})$, should be a random sample from an exponential distribution of unknown mean.

Jaquet et al. (2000) proposed a temporal occurrence model using the Cox process. This is a doubly stochastic process with

$$\Pr(N(s, s+t) = k) = \mathbb{E} \left(\frac{Z(s, s+t)^k}{k!} e^{-Z(s, s+t)} \right), \quad (6)$$

where

$$Z(s, s+t) = \int_s^{s+t} \omega(t) dt \quad (7)$$

and $\omega(t)$ is the conditional random frequency of events. The process of determining $\omega(t)$ is not spelled out in detail, involving geostatistical methods, simulation, variograms, and a conditioning technique. The advantage of Eq. 6 over the Poisson process is that it introduces the possibility of correlation in time.

Renewal Processes

The homogeneous Poisson process without trend is a special case of a *renewal process*. In a renewal process, the intervals between events are independent and identically distributed, and so

$$\lambda(t) = \frac{f(t-s; \boldsymbol{\theta})}{1 - F(t-s; \boldsymbol{\theta})}, \quad t > s, \quad (8)$$

where the most recent event occurred at time $s < t$ and $f = F'$ is a density with parameter vector $\boldsymbol{\theta}$. This was first suggested for eruption onsets by Wickman (1966a), who coined the term *age-specific eruption rate* for Eq. 8. The character of the renewal process is that only the elapsed time since the last eruption controls the time to the next eruption. Previous eruptions exert an influence only through their contribution to the parameter estimates $\hat{\boldsymbol{\theta}}$. A number of tests to check whether a series of eruptions is consistent with a renewal process can be found in

Bebbington and Lai (1996a, b), Ogata (1988), Reymont (1969), and Bebbington (2013); one checks for lack of trend, independence of successive repose lengths, and fit to a particular distribution (the exponential in the special case of the Poisson process).

Given a density $f(u; \theta)$ and observed inter-onset times r_i , $i = 1, \dots, n-1$, the parameters θ can be estimated by maximum likelihood. That is, the values are chosen, either algebraically or numerically, to maximize the likelihood

$$L(r_1, \dots, r_{n-1}, t-s; \theta) = [1 - F(t-s; \theta)] \prod_{i=1}^{n-1} f(r_i; \theta), \quad (9)$$

where the first term accounts for the current unfinished interval. A similar term for the incompletely observed interval before the first recorded onset is omitted, as the beginning of the observation window is usually unknown. Occasionally other methods are used. The method of moments involves setting the parameter values to equalize the observed and theoretical moments. As many moments are used as one has parameters. For example, two parameters can be fitted using the mean and standard deviation. An alternative is based on minimizing the sum of squared differences between an observed curve and that produced by the model. This is usually done using linear regression, but direct numerical optimization is possible.

If f is the exponential density $f(u) = ve^{-vu}$, then

$$\lambda(t) = \frac{ve^{-v(t-s)}}{1 - (1 - e^{-v(t-s)})} = v, \quad (10)$$

and we recover the homogeneous Poisson process. Klein (1982) tested the eruptive patterns of Hawaiian volcanoes against a Poisson process (random model) using this formulation and showed that repose following large eruptions differed significantly from the whole. Eq. 10 is sometimes referred to as the memoryless property, as it says that the time elapsed since the last eruption provides no information about the time of the next eruption.

Thorlaksson (1967) suggested that an over-dispersed (standard deviation σ greater than the mean μ) sequence of inter-onset intervals should be modeled by the Pareto density $f(u) = a(1 + bu)^{-a/b - 1}$, $u > 0$, with intensity

$$\lambda(t) = \frac{a}{1 + b(t-s)}, \quad t > s, \quad (11)$$

where $\hat{a} = 2\sigma^2/(\mu^3 + \mu\sigma^2)$ and $\hat{b} = (\sigma^2 - \mu^2)/(\mu^3 + \mu\sigma^2)$ are estimated by the method of moments. This models a decrease in eruption probability with time and was applied to Colima by Medina Martinez (1983). For under-dispersed inter-onset intervals, Thorlaksson (1967) suggested the density

$$f(u) = \begin{cases} (a - bu) \exp(-au - (b/2)u^2) \sqrt{4/\pi - 1} < \sigma/\mu < 1, \\ b(u - a) \exp(-(b/2)(u - a)^2) \sigma/\mu < \sqrt{4/\pi - 1} u > 0, \end{cases} \quad (12)$$

leading to the intensity

$$\lambda(t) = \begin{cases} a + b(t-s) & \sqrt{4/\pi - 1} < \sigma/\mu < 1 \\ b(t-s-a) & H(t-s-a)\sigma/\mu < \sqrt{4/\pi - 1} t > s, \end{cases} \quad (13)$$

where

$$H(x) = \begin{cases} 1 & x > 0 \\ 0 & x \leq 0 \end{cases} \quad (14)$$

is the Heaviside function. These imply an increase in eruption probability with time, while the case $\sigma/\mu < \sqrt{4/\pi - 1}$ adds a loading time a , and the parameters can be estimated as $\hat{a} = \mu - \sigma/\sqrt{4/\pi - 1}$ and $\hat{b} = (2 - \pi/2)/\sigma^2$. This latter case was applied to the volcanoes Hekla and Katla, which had ratios σ/μ of 0.44 and 0.43, respectively. Of the other 28 volcanoes examined, none had a ratio less than 0.76, and 23 had a ratio of greater than unity.

Settle and McGetchin (1980) fitted a Gaussian density to the inter-onset times of eruptions at Stromboli over four days. Note that the formulation (8) is always strictly positive, even for distributions such as the Gaussian whose support includes the negative axis. However, fitting and

simulating such processes presents certain practical difficulties.

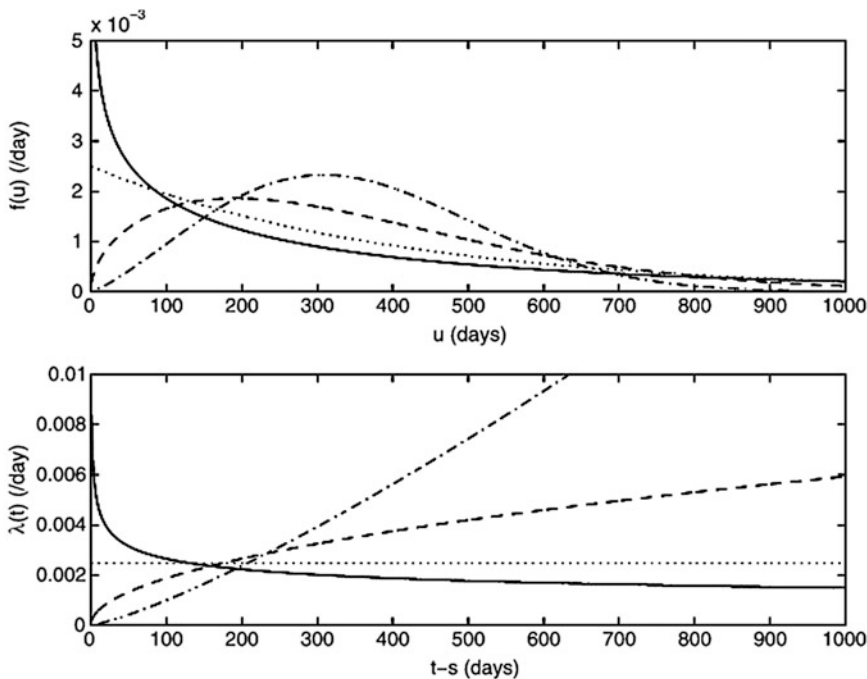
Bebbington and Lai (1996a) fitted renewal models to a number of volcanoes, using the Weibull density with scale parameter β and shape parameter α :

$$f(u) = \alpha\beta(\beta u)^{\alpha-1} \exp(-(\beta u)^\alpha), \quad u > 0. \quad (15)$$

The intensity is then

$$\lambda(t) = \alpha\beta(\beta(t-s))^{\alpha-1}, \quad t > s. \quad (16)$$

We see that while $\alpha = 1$ is the exponential distribution, $\alpha < 1$ corresponds to an “over-dispersed,” or clustering, distribution (cf. Crandell et al. 1975) and $\alpha > 1$ to a more periodic distribution with a mode at $u = \beta^{-1}(1 - 1/\alpha)^{1/\alpha}$. The density and intensity are shown in Fig. 4. Watt et al. (2007) showed that the model applied also at a much finer temporal scale to Vulcanian eruption sequences.



Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 4 Density f and conditional intensity λ for the Weibull distribution with $\beta = 0.0025$ and $\alpha = 0.75$

(solid line), $\alpha = 1$ (dotted line), $\alpha = 1.5$ (dashed line), and $\alpha = 2.25$ (dot-dash line)

De la Cruz-Reyna and Carrasco-Nunez (2002) suggested using the gamma density, which also allows for clustering and periodicity, but the computational details are more complex. Also, the Weibull is the consequence of a material failure model (Voight 1988), and in any case, little qualitative difference is usually observed between Weibull and gamma distributions for small samples, such as eruptive records. Note that the intensity (Eq. 16) is monotonic and can model either increasing probability of eruption as the repose time increases, such as at Hekla (Wickman 1966a), or decreasing probability with increasing repose time, such as at Colima (Medina Martinez 1983). Bebbington and Lai (1996a, b) provided a number of tests for assessing the applicability of the Weibull distribution, based on the fact that r_i^β , where $r_i = t_{i+1} - t_i$, should be exponentially distributed.

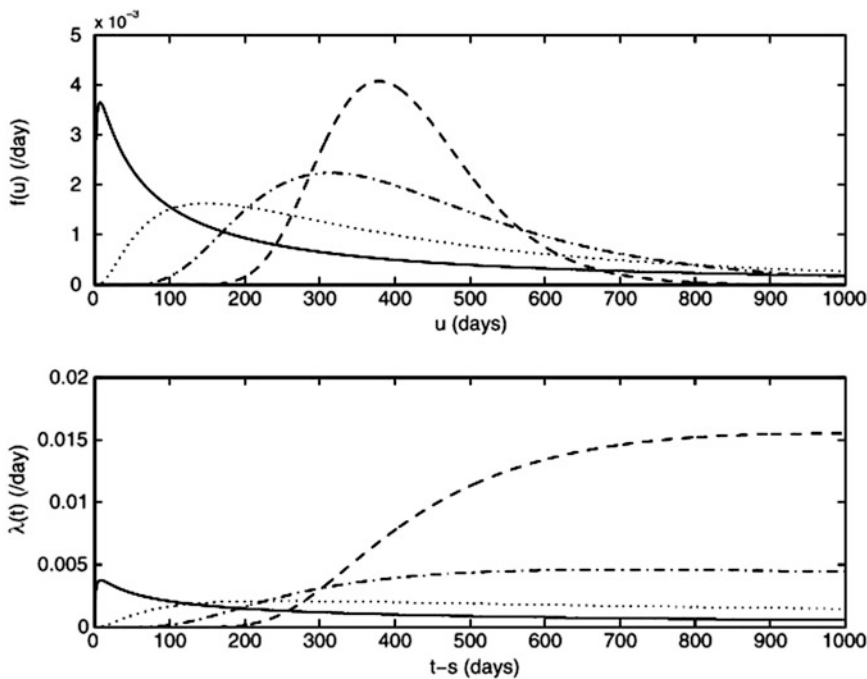
The lognormal density,

$$f(u) = \frac{1}{u\sigma\sqrt{2\pi}} \exp \left[-\frac{(\ln u - \mu)^2}{2\sigma^2} \right], \quad u > 0, \quad (17)$$

with intensity

$$\lambda(t) = \frac{\frac{\sqrt{2}}{(t-s)\sigma\sqrt{\pi}} \exp \left[-\frac{(\ln(t-s)-\mu)^2}{2\sigma^2} \right]}{\operatorname{erfc} \left(\frac{\ln(t-s)-\mu}{\sqrt{2}\sigma} \right)}, \quad t > s, \quad (18)$$

was also considered by Bebbington and Lai (1996a) and used by Marzocchi and Zaccarelli (2006) and Eliasson et al. (2006). As the density (Eq. 17) is unimodal at $u = \exp(\mu - \sigma^2)$, this works better for non-clustering events, such as the larger eruptions considered in Marzocchi and Zaccarelli (2006). Eliasson et al. (2006) noted significant departure from the distribution in one of their models. The density and intensity are shown in Fig. 5. Note that Eq. 18 is “upside-



Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 5 Density f and conditional intensity λ for the lognormal distribution with $\mu = 6$ and $\sigma = 2$ (solid

line), $\sigma = 1$ (dotted line), $\sigma = 0.5$ (dashed line), and $\sigma = 0.25$ (dot-dash line)

down bathtub shaped,” i.e., that as t increases, it rises to a maximum and then declines.

The log-logistic density,

$$f(u) = \frac{\eta^\gamma u^{\gamma-1}}{(1 + \eta u^\gamma)^2}, \quad u > 0, \quad (19)$$

with a mode at $u = \max\{0, (\gamma - 1)/((\gamma + 1)\eta)^{1/\gamma}\}$ was used by Connor et al. (2003) to model the intervals between Vulcanian explosions of Soufriere Hills volcano, Montserrat. The intensity

$$\lambda(t) = \frac{\eta^\gamma (t - s)^{\gamma-1}}{1 + \eta (t - s)^\gamma}, \quad t > s, \quad (20)$$

has a maximum at $t = s + \max\{0, (\gamma - 1)/\eta^{1/\gamma}\}$, prior to which it is increasing and subsequently decreasing. Thus, we can see that the log-logistic distribution possesses some of the character of the lognormal distribution, as does the inverse-Gaussian density

$$f(u) = \sqrt{\frac{\mu}{2\pi\alpha^2 u^3}} \exp\left[-\frac{(u - \mu)^2}{2\alpha^2 \mu u}\right], \quad \mu, \alpha > 0, \quad (21)$$

(sometimes called the Brownian Passage Time distribution) examined in Garcia-Aristizabal et al. (2012).

Pyle (1998) used the power-law density

$$f(u|u > c) = \frac{u^{-1-1/b}}{bc^{-1/b}}, \quad u > c > 0, \quad (22)$$

to model the tail of the inter-onset distribution. Note that the density is not defined for $c = 0$. The intensity is

$$\lambda(t) = \frac{1}{b(t - s)}, \quad t > s, \quad (23)$$

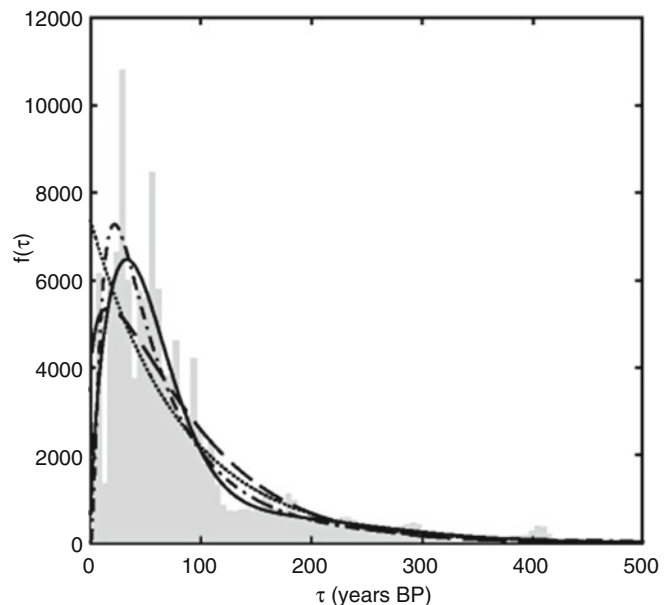
from which we see that the parameter b controls the rate of decay of the intensity or, equivalently, the “thickness of the tail.”

Using event data derived from Monte Carlo simulation of a stratigraphic profile, Turner et al. (2008a) fitted a renewal density consisting of a mixture of Weibull densities, as exemplified in Fig. 6, to model a multimodal repose distribution. Cronin et al. (2001) proposed a variant on this, where onsets occurred in “episodes,” each of variable numbers of eruptions, with different repose distributions corresponding to inter- and intra-episode repose.

Applications of renewal models to describe volcano onset records are summarized in Marzocchi and Bebbington (2012). The drawback of renewal models is that they commonly fail to explain variations in eruption rate, corresponding to changes in

Volcanic Eruptions: Stochastic Models of Occurrence Patterns,

Fig. 6 Histogram of 122,000 sampled inter-event times based on 1,000 Monte Carlo runs. Curves show the fitted densities for this data set: dotted line exponential distribution, dot-dash lognormal, dashed Weibull, solid mixture of Weibull (Bull Volcanol 70: 513, Fig. 4, © Springer-Verlag 2007)



activity level, although Wickman (1966a) did suggest using a small number of different levels. This idea was taken up in Bebbington (2007), where a renewal process with parameters controlled by a hidden Markov model was fitted to flank eruptions from Mt Etna. Using the Weibull distribution (Eq. 15) for the inter-onset times, the preferred model had long sequences of approximately Poisson behavior, interspersed with shorter sequences of more frequent and regular eruptions.

Markov Processes

A *Markov process* (continuous time Markov chain) $Z(t)$, $t \geq 0$, has transition intensity matrix $Q = (q_{ij})$, where

$$q_{ij} = \lim_{t \downarrow 0} \frac{\Pr(Z(s+t) = j | Z(s) = i)}{t} \geq 0, \quad (24)$$

for all $i \neq j$ and $-q_i = q_{ii} = -\sum_{j \neq i} q_{ij}$. Thus, the sojourn time in state i is exponential with mean $1/q_i$, and when it ends, the process moves to state j with probability $p_{ij} = q_{ij} / \sum_{k \neq i} q_{ik}$.

Wickman (1966b) proposed the use of Markov processes to model eruption-repose patterns. This entailed defining a number of eruptive and repose states, with prescribed permitted transitions between them. The initial models (Wickman 1966b) allowed for one or two magma chambers, with eruptions and “solidification” of the vent occurring in Poisson processes of differing rate. In the two-state case, this solidification can only be removed by an eruption from the second, larger and deeper, magma chamber. If the two magma chambers erupt at rates ν and μ ($\nu > \mu$), respectively, and the vent solidifies at rate η , then the eruption rate can be derived as

$$\lambda(t) = \frac{\mu\eta + \nu(\mu + \nu + \eta) \exp[-(\nu + \eta)(t - s)]}{\eta + \nu \exp[-(\eta - \nu)(t - s)]}, \quad t > s, \quad (25)$$

where the previous eruption occurred at time s . The parameters can be estimated by numerically fitting this equation to the empirical renewal intensity.

Wickman (1976) extended this to six possibilities modeled on various well-known volcano types and simulated them to produce reasonable appearing synthetic data. The process of fitting these models to observed data has required manual identification of the various states in the data (Carta et al. 1981; Martin and Rose 1981), but the use of hidden Markov model techniques to automate the process has been suggested (Bebbington 2007). In a Markov process, the time between onsets is the passage time through the eruptive and repose states, which will have a compound exponential distribution, i.e., be the sum of a number of exponential random variables of differing means.

Time- and Size-Predictable Models

Bacon (1982) noted that the intervals between rhyolitic and basaltic eruptions in the Coso Range (California) appeared to be proportional to the volume of the preceding eruption. This *time-predictable* behavior (cf. Shimazaki and Nakata 1980) is most simply explained by assuming that the rate of magma input is constant and an eruption occurs when a certain magma level is reached. However, Bacon (1982) suggested the covariate of accumulated extensional strain as the controlling variable and considered eruptions a passive response of the magmatic system to the tectonic stress field.

Similar behavior was observed for Fuego volcano, Guatemala (Martin and Rose 1981), together with a weaker proportionality between volume and the length of repose before an eruption, although the authors did note that “the volume estimates probably have an error of less than ± 1 order of magnitude.” Santacroce (1983) found a similar phenomenon at the Somma-Vesuvius complex and suggested that eruptions there occur in cycles, of approximately 3,000 years, each initiated by a large-scale Plinian eruption and terminated by a long repose of several centuries. Klein (1982) found evidence of time-predictable behavior at both Kilauea and Mauna Loa (Hawaii).

Wadge and Guest (1981) suggested that the time-predictable model has a limited forecasting

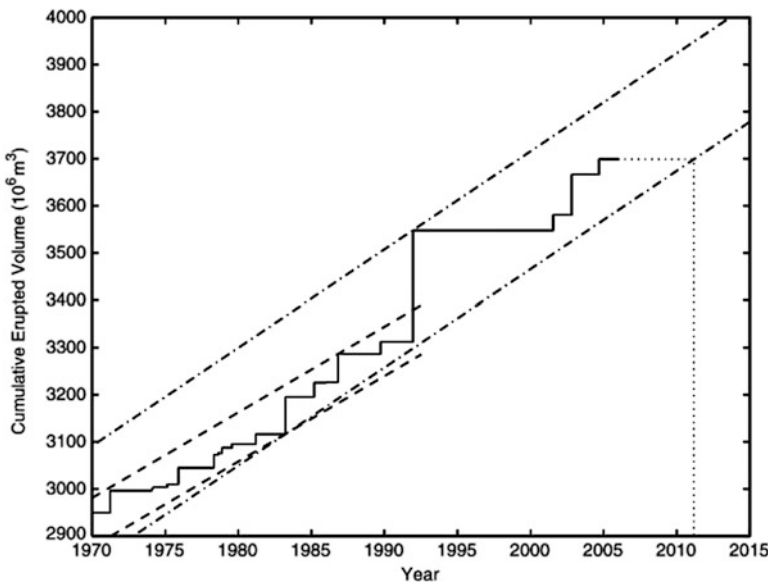
ability, based on drawing an empirical envelope on the recent eruptive volume curve. This has a slope determined as the mean eruptive volume rate and limits defined by the extreme points of the curve in the considered history. This enables a forecast of both the longest the current repose is likely to last and the maximum volume of the next eruption, provided that steady state is maintained. Figure 7 (from the first edition of this work) shows an example using flank eruptions from Mt Etna 1970–2006. The next flank eruption occurred on August 13, 2008, well within the predicted envelope. The lava volume was small, but the relatively large lava volume ($4.5 \times 10^6 \text{ m}^3$) of the preceding summit eruption at the Southeast Crater was also well within limits. A similar forecast was made for Cerro Negro (Hill et al. 1998), and the method has been adapted to the forecasting of sector collapses (Zernack et al. 2012). Wadge (1982) estimated, however, that volcanoes may spend only 1/4–1/2 of the time in steady state, being dormant much of the remainder. Changes

in eruptive rate have also been found in a number of historical records (Bebbington 2010; Burt et al. 1994; De la Cruz-Reyna 1991; Mulargia et al. 1987; Smethurst et al. 2009; Turner et al. 2011).

De la Cruz-Reyna (1991) proposed a general *load-and-discharge* model, where the “energy” of the volcano increases at a constant rate σ between eruptions. The i th eruption occurs when this exceeds the threshold H_i , during which the stored energy drops to the threshold L_i , releasing an energy $\xi_i = H_i - L_i$. The interval between eruptions is thus

$$r_i = t_{i+1} - t_i = \frac{H_{i+1} - H_i + \xi_i}{\sigma} \quad (26)$$

Hence, successive repose intervals are not independent, as they include a common threshold term, and should be negatively correlated. In general, the thresholds can be random, but this formulation includes the time-predictable model. In this, the thresholds H_i are constant, and thus the interval



Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 7 Volume-time curve for Mt Etna flank eruptions (solid line). Dashed limits based on 1970–1990 data were exceeded, or equivalently the volcano departed from steady-state, by the 1991 eruption. Dot-dash limits based on 1970–2006 data have a steeper slope and wider

limits to accommodate the 1991 eruption. The dotted lines show that the method predicts the next eruption to occur by the end of February 2011 and the eruptive volume to be less than $142.1 + 20.8 t$ (10^6 m^3), where t is the number of years to the date of the eruption from January 2007

$$r_i = \xi_i / \sigma \quad (27)$$

is proportional to the energy of the preceding eruption, and successive intervals are independent. Alternatively, if the thresholds L_i are held constant ($=L$, say), $H_i = \xi_i - L$, and from Eq. 27,

$$r_i = (H_{i+1} - L) / \sigma. \quad (28)$$

Thus, successive intervals are again uncorrelated, and the energy drop of the eruption is proportional to the length of the interval preceding it. This is termed the *size-predictable* model. The formulation (26) was placed in an estimable framework (Bebbington 2008) using the intensity

$$\lambda(t) = \exp \{ \alpha + v[\rho t - V(t)] \}, \quad (29)$$

where $V(t) = \sum_{k:t_k < t} v_k$ is the cumulative volume erupted prior to time t and α , $v > 0$, and $\rho > 0$ are parameters to be estimated.

In a global analysis of eruptions, De la Cruz-Reyna (1991) noted the existence of a regression relationship $\log \lambda_0 = 3.494 - 0.789 \text{ VEI}$ with correlation -0.999 , where λ_0 is the Poisson intensity for eruptions of a given VEI. This led De la Cruz-Reyna (1993) (see also De la Cruz-Reyna and Carrasco-Nunez 2002) to model the onsets of different VEI events as independent Poisson processes, in a special case of the size-predictable model. This approach was also used in the quantification of volcanic risk due to Vesuvius by Scandone et al. (1993b), based on the sequence of equilibrium states (repose, persistent activity, intermediate eruption, final eruption) identified by Carta et al. (1981).

Burt et al. (1994), in an examination of the basaltic volcano Nyamuragira (Zaire), formulated the “pressure-cooker” and “water-butt” models. The former, where an eruption occurs when the magma volume reaches a threshold determined by the rock strength and reservoir shape, is the time-predictable model, while the latter, where eruptions always drain the magma reservoir to the same level, is the size-predictable model. Burt et al. (1994) suggested physical explanations for

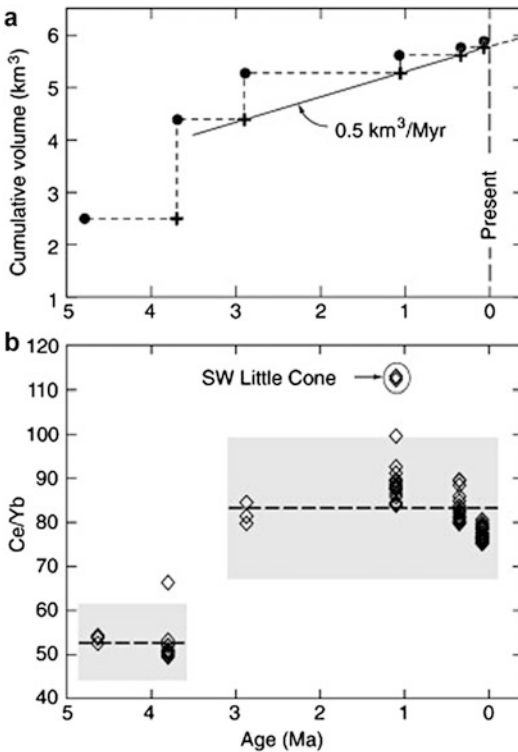
stochastic perturbations observed in these models and proposed that a volcano could be tested for time predictability by performing a regression analysis of (r_i) on $(v_i^{(1)})$, where $r_i = t_{i+1} - t_i$ and $t_i, v_i^{(1)}$ are the onset time and eruptive volume of the i th eruption. Similarly the size-predictable model entails a regression analysis of $(v_i^{(2)})$ on (r_i) , where $v_i^{(2)} = v_{i+1}^{(1)}$. Sandri et al. (2005) generalized the test for time predictability to a regression analysis of $(\log v_i^{(1)})$ on $(\log r_i)$, so that an estimated slope of b significantly greater than zero implies a time-predictable relation $u_i^{(1)} \propto r_i^b$. Marzocchi and Zaccarelli (2006) performed a similar regression analysis of $(\log v_i^{(2)})$ on $(\log r_i)$ for the size-predictable model, resulting in a relation $u_i^{(2)} \propto r_i^b$. Because both the repose and volume distributions are highly skewed, the logarithmic transformation is advisable to reduce the high leverage of the tail points. Valentine and Perry (2007) analyzed eruption onsets in a basaltic volcanic field using the time- and size-predictable models, interpreting the former as showing tectonic control (Fig. 8).

Marzocchi and Zaccarelli (2006) proposed a renewal model based on the time-predictable model. In their *open conduit system* model, the probability density of the time u to the next eruption is conditional on the previous eruptive volume v . Using a lognormal distribution, this is

$$f(u | v) = \frac{1}{u\sigma\sqrt{2\pi}} \exp \left[-\frac{(\log u - (a + bv))^2}{2\sigma^2} \right] \quad (30)$$

where a, b are the intercept and slope regression coefficients from the regression of $(\log v_i^{(1)})$ on $(\log r_i)$. In maximum likelihood estimation, this can be shown (Marzocchi and Bebbington 2012) to be equivalent to a lognormal renewal model with mean repose proportional to v^b , where v is the volume of the preceding eruption. A Bayesian hierarchical framework incorporating (Eq. 30) has been developed by Passarelli et al. (2010a).

Marzocchi et al. (2004b) estimated the VEI of the next eruption of Mt Vesuvius by considering



Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 8 (a) Plot of volume-age relationships in Plio-Pleistocene basalts of Southwestern Nevada Volcanic Field. Filled circles are cumulative volume versus age of each episode, while crosses show time-predictable trend of cumulative volume just prior to an episode versus age of the episode. (b) Ce/Yb versus age for Plio-Pleistocene basalts. The shift to higher but variable Ce/Yb beginning at 2.9 Ma indicates a shift to generally lower degrees of partial melting. SW Little Cone is a very small volcano with nepheline-normative composition, suggesting smaller degree of melting at higher pressure than other basalts in the 1.1 Ma episode. Dashed lines and shaded fields represent the mean and two standard deviations, respectively, for Ce/Yb values of >3 Ma and <3 Ma basalts (Reprinted from Valentine GA, Perry FV (2007) Tectonically controlled, time-predictable basaltic volcanism from a lithospheric mantle source (central Basin and Range Province, USA). *Earth Planet Sci Lett* 261:211 © 2007 Elsevier B.V., with permission from Elsevier)

(a) the global catalog, (b) a catalog of “analog” volcanoes, and (c) the catalog of Vesuvius itself. In each case, the catalogs are trimmed to those eruptions following a repose interval of at least the presently observed 60 years. The estimated probability of an eruption of $\text{VEI} \geq 5$ ranged from 1 %

to 20 % and could not be considered negligible. This has implications for the Emergency Plan of Mount Vesuvius, which assumes a maximal $\text{VEI} = 4$.

Orsi et al. (2011) analyzed the volume record of the Campi Flegrei, but the hazard model (Selva et al. 2010) implicitly assumes the volume to be independent of repose length, an approach also taken in Bebbington (2008). Thus far no working model for the size of the next eruption has been developed. Some volcanoes appear to have power-law (Bebbington 2008; Luongo and Mazzarella 2001) or Pareto (Mendoza-Rosas and De La Cruz-Reyna 2008) size distributions mimicking earthquake magnitude statistics and arguing against size predictability. Other volcanoes with less well-developed records have a more regular pattern of VEIs, but the VEI is too crude a measure of “size” to draw any conclusions from this.

Chaos and Fractals

A sufficiently complicated deterministic model is indistinguishable by finite observation from a random process. However, it is possible to construct systems with few degrees of freedom but very complex dynamics, which are known as *chaotic systems*, or *strange attractors*. This possibility can be tested by examining the correlation properties of the observation sequence. The underlying principle of fractals is one of *self-similarity*, i.e., that the statistical properties of the process are invariant at differing timescales (or space scales).

Sornette et al. (1991) computed the dimension D of the attractor as a function of the dimension d of the embedding phase space. Given onset times t_i , let $X_i^{(k)} = r_{i+k-1}$ for $k = 1, \dots, d-1$, where $r_i = t_{i+1} - t_i$, for $i = 1, \dots, n-1$. Then the correlation function is determined as

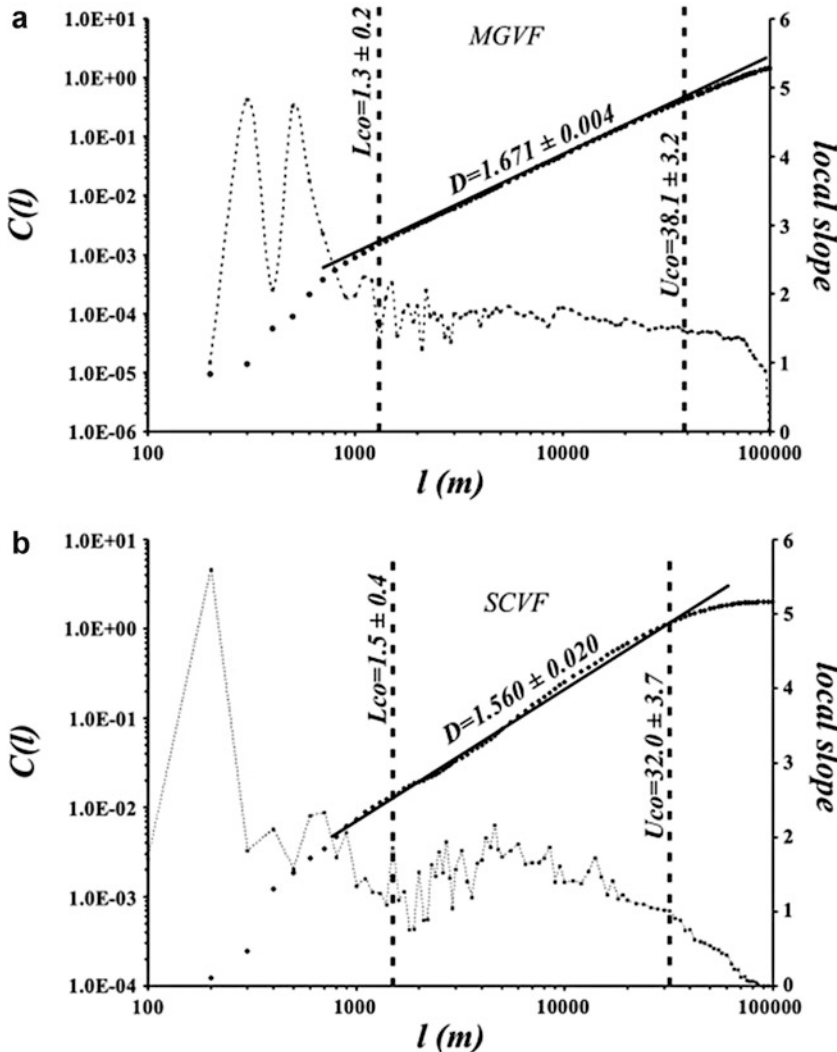
$$C_d(w) = \frac{1}{n^2} \# \{i, j : \|\mathbf{X}_i - \mathbf{X}_j\| < w\}, \quad (31)$$

where $\mathbf{X}_i$ is the vector $(X_i^{(1)} \dots X_i^{(d)})$. The correlation dimension is then defined by the power-law relation

$$C(w) \sim w^{D(d)}. \quad (32)$$

If this saturates at some value $D_{\max}$ as d increases, this is taken as evidence that the underlying dynamics is deterministic and has dimension of the order of $D_{\max}$. Applications to the volcanoes Piton de la Fournaise (La Reunion Island) and Mauna Loa/Kilauea (Hawaii) produced estimates of $D_{\max}$ of 1.7 and 4.6, respectively. The latter

value was interpreted as evidence of two independent dynamical systems for the two Hawaiian volcanoes, contradicting the conclusion of Klein (1982). Sornette et al. (1991) also constructed a series of “return maps” (r_{i+1} as a function of r_i) by truncating the data in various ways. A similar two-point correlation analysis was used by Mazzarini et al. (2010) to describe the spatial vent clustering in the Trans-Mexican Volcanic Belt (Fig. 9).



Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 9 The correlation integral $C(l)$ versus l in a logarithmic plot. D is the fractal exponent of the power law that describes the self-similarity of vent distribution, (a) MGVE, (b) SCVF. Lco lower cutoff, Uco upper cutoff (Reprinted from Mazzarini F, Ferrari L, Isola I (2010)

Self-similar clustering of cinder cones and crust thickness in the Michoacan-Guanajuato and Sierra de Chichinautzin volcanic fields, Trans-Mexican Volcanic Belt. Tectonophysics 486:62 © 2010 Elsevier B.V., with permission from Elsevier)

Dubois and Cheminee (1991) used the Cantor dust model for volcanic eruption onsets. In this method, the eruptive record is divided into segments of length T , and the fraction of these intervals containing at least one onset is denoted X . If fractal clustering with fractal dimension D is present, then $X \sim T^{1-D}$. This implies a renewal model with intensity $\lambda(t) = D/(t-s)$, $t > s$, where the last eruption occurred at time s (cf. Eq. 23). An obvious method for estimating D is a regression analysis of $(\log X_i)$ against $(\log T_i)$ for various interval lengths T_i . Applied to the Piton de la Fournaise (La Reunion Island), two slopes were observed, one with $D \approx 0.5$ for $T < 10$ months and one with $D \approx 0.7$ for $T > 10$ months. Kilauea had two regimes with $D = 0.58$ for $T < 24$ months and $D = 0.81$ for $T > 24$ months, while Mauna Loa had $D \approx 0.35$ for $T < 6$ years and $D \approx 0.85$ for $T > 6$ years. The two regimes all indicated clustering at short repose lengths, and more regular activity for eruptions separated by longer repose. This was interpreted as evidence of multiple magma chambers, with different sizes and recharge rates, at different depths. Applied to a combined catalog of flank and summit eruptions of Mt Etna 1950–1987, the result was a single regime with $D = 0.88$. This high value of D was interpreted as being characteristic of the regular eruptive activity of the volcano.

Marzocchi (1996) calculated the statistic

$$a_{\max} = \max_{1 \leq s \leq 5} a(s), \quad (33)$$

where $a(s)$ is defined in Eq. 3, although a process might first be de-trended by subtracting an autoregressive model. A value $a_{\max} > a_c^*$ indicates a chaotic system, while a value $a_{\max} < a_s^*$ indicates a stochastic system. Intermediate values provide insufficient evidence for either. The critical values a_c^* and a_s^* are determined by simulation. Both Vesuvius and Etna evidenced significantly stochastic behavior.

A multifractal analysis using the Grassberger-Procaccia algorithm was performed by Godano and Civetta (1996). The correlation interval is calculated as

$$C(t, q) = \frac{1}{n-1} \sum_{i=1}^{n-1} \left[\frac{1}{n - 2 \sum_{j \neq i} H(t - |r_i - r_j|)} \right]^q, \quad (34)$$

where $r_1, \dots, r_{n-1}$ are the inter-onset times and $H(\cdot)$ is the Heaviside function. The *generalized dimensions* are then evaluated as

$$D_{q+1} = \phi(q)/q, \quad (35)$$

where $\phi(q)$ is the estimated slope from a regression analysis of $\log C(t, q)$ versus $\log t$. The fractal, information, and correlation dimensions are D_0 , D_1 , and D_2 , respectively. More generally, D_{q+1} is constant as q varies for a homogeneous fractal distribution but varies with q for a multifractal distribution. The latter implies that local clustering properties are different from global clustering properties. Applied to eruptions of Vesuvius, the estimated dimensions were $D_0 = 1.05 \pm 0.07$, $D_2 = 0.75 \pm 0.04 \approx D_\infty = 0.73 \pm 0.02$, and $D_{-\infty} = 1.09 \pm 0.13$. This indicates that at a global scale, the catalog exhibits Poisson behavior ($D = 1$), and the most intense clustering is not a particularly strong effect. The catalog exhibits only weak multifractal behavior, and this is only for $q < 1$. Thus, the cyclic behavior outlined by Carta et al. (1981) cannot be identified statistically, as the intervals between events in a cycle are not distinguishable from those between cycles.

Nishi et al. (2001) examined the long-term memory of the time series of 3,796 eruptions of Sakurajima 1981–1999 using the Hurst exponent. Let $r_i = t_{i+1} - t_i$, $i = 1, \dots, n-1$, be the inter-onset times, and set

$$X(k, t, m) = \sum_{i=1}^k \left(r_{i+t-1} - \frac{1}{m} \sum_{j=1}^m r_{j+t-1} \right). \quad (36)$$

Then the *self-adjust range* is

$$R(t, m) = \max_{1 \leq k \leq t} X(k, t, m) - \min_{1 \leq k \leq t} X(k, t, m), \quad (37)$$

which is rescaled by the standard deviation

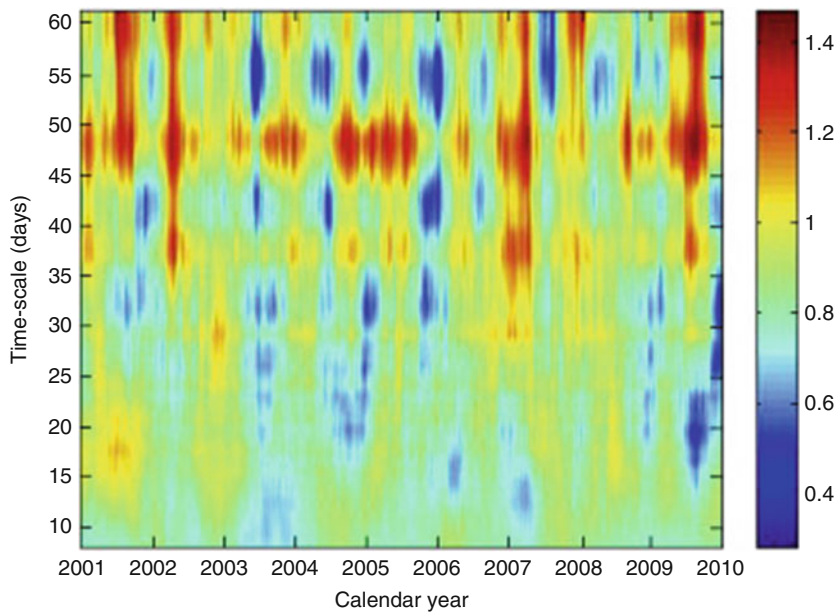
$$S(t, m) = \sqrt{\frac{1}{t} \sum_{i=1}^t \left(r_{i+m-1} - \frac{1}{m} \sum_{j=1}^m r_{j+t-1} \right)^2}, \quad (38)$$

and averaging over the length of the time series, the *Hurst exponent* H is then defined by

$$\frac{\sum_{m=1}^{n-1} R(t, m)}{\sum_{m=1}^{n-1} S(t, m)} = \left(\frac{t}{2} \right)^H. \quad (39)$$

The value of H can be estimated by a regression analysis of Eq. 39, after taking logarithms. The Poisson process produces a value of $H = 0.5$. A value of H greater than this implies persistent behavior, with each value depending on the previous

one. Anti-persistence behavior is characterized by $H < 0.5$. The Hurst exponents for Sakurajima were $H = 0.72$ for the whole series and $H = 0.74$ for a “high-frequency” period 1983–1985. A Monte Carlo test consisting of randomly reordering the observed inter-onset times and recalculating H confirms that these values are significantly greater than 0.5. A similar exercise was conducted on 52,691 “exhalations” of Popocatepetl volcano 2000–2009 (Alvarez-Ramirez et al. 2011), with $H = 0.847$. This was extended to a scale-dependent Hurst exponent, which exhibited peaks in autocorrelation from 40 to 60 days, and an apparent quasiperiodic cycle (Fig. 10). Telesca and Lapenna (2005) performed the same analysis for eruptions of $\text{VEI} \geq 0$ of 14 volcanoes worldwide. The requirements were a minimum of 40 events during 1800–2000. Seven of the volcanoes, with Hurst exponents ranging from 0.79 to 1.3, were characterized by persistent behavior, while the remainder, with Hurst exponents ranging from 0.60 to 0.73, were not significantly different from a Poisson process. Bebbington (2010) showed that this was better explained by



Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 10 Hurst exponent as a function of physical time and timescale. A regular structure is observed for timescales about 4,050 days (Reprinted from Alvarez-Ramirez J, Sosa E, Hernandez-Martinez E (2011)

In-phase dynamics of the exhalation sequence in Popocatepetl volcano and slow-slip events in Cocos-North American plate boundary. J Volcanol Geotherm Res 200:86 © 2010 Elsevier B.V., with permission from Elsevier)

variation in the intensity rather than dependence between onsets.

Telesca et al. (2002) examined 35 sequences of $\text{VEI} \geq 0$ eruptions (minimum 25 eruptions 1800–2000) using the Fano factor method. The Fano factor is defined as the variance of the number of events in a specified interval width Δ divided by the mean,

$$\text{FF}(\Delta) = \frac{(\Delta/T) \sum_{i=1}^{T/\Delta} N_i^2 - \left((\Delta/T) \sum_{i=1}^{T/\Delta} N_i \right)^2}{(\Delta/T) \sum_{i=1}^{T/\Delta} N_i}, \quad (40)$$

where $N_i = N((i-1)\Delta, i\Delta)$. Telesca et al. (2002) represented the FF of a fractal point process as a monotonic power law

$$\text{FF}(\Delta) = 1 + \left(\frac{\Delta}{\Delta_0} \right)^\alpha, \quad (41)$$

for $\Delta > \Delta_0$, where Δ_0 is the fractal onset time, marking the lower limit for significant scaling behavior, with negligible clustering below this. Again the fractal exponent α is estimated by a regression analysis of Eq. 41, after taking logarithms. For Poisson processes, the FF is approximately one for all interval lengths, and so $\alpha \approx 0$. Of the 35 sequences examined, 30 had fractal exponents ranging from 0.2 to 0.9 (mean of 0.5). The remaining five exhibited no value Δ_0 above which Eq. 41 held.

Gusev et al. (2003) examined clustering of eruptions on Kamchatka during the last 10,000 years using the Weibull density (Eq. 15) for the inter-onset times, the correlation dimension (Eq. 32), and the Hurst exponent (Eq. 39). No self-similar behavior was found for intervals greater than 800 years, but for shorter delays, there was a significant degree of clustering. Gusev et al. (2003) also generalized the correlation dimension by replacing (Eq. 31) by a weighted version, where the contribution of a pair of events is proportional to the product of their eruptive volumes. This identified self-similar behavior over the entire range of inter-onset times (100–10,000 years). Extended to the global catalog (Siebert and Simkin 2002),

Gusev (2008) showed that larger events are over-represented in periods of higher activity (Fig. 11), which runs counter to what would be expected from incomplete observation.

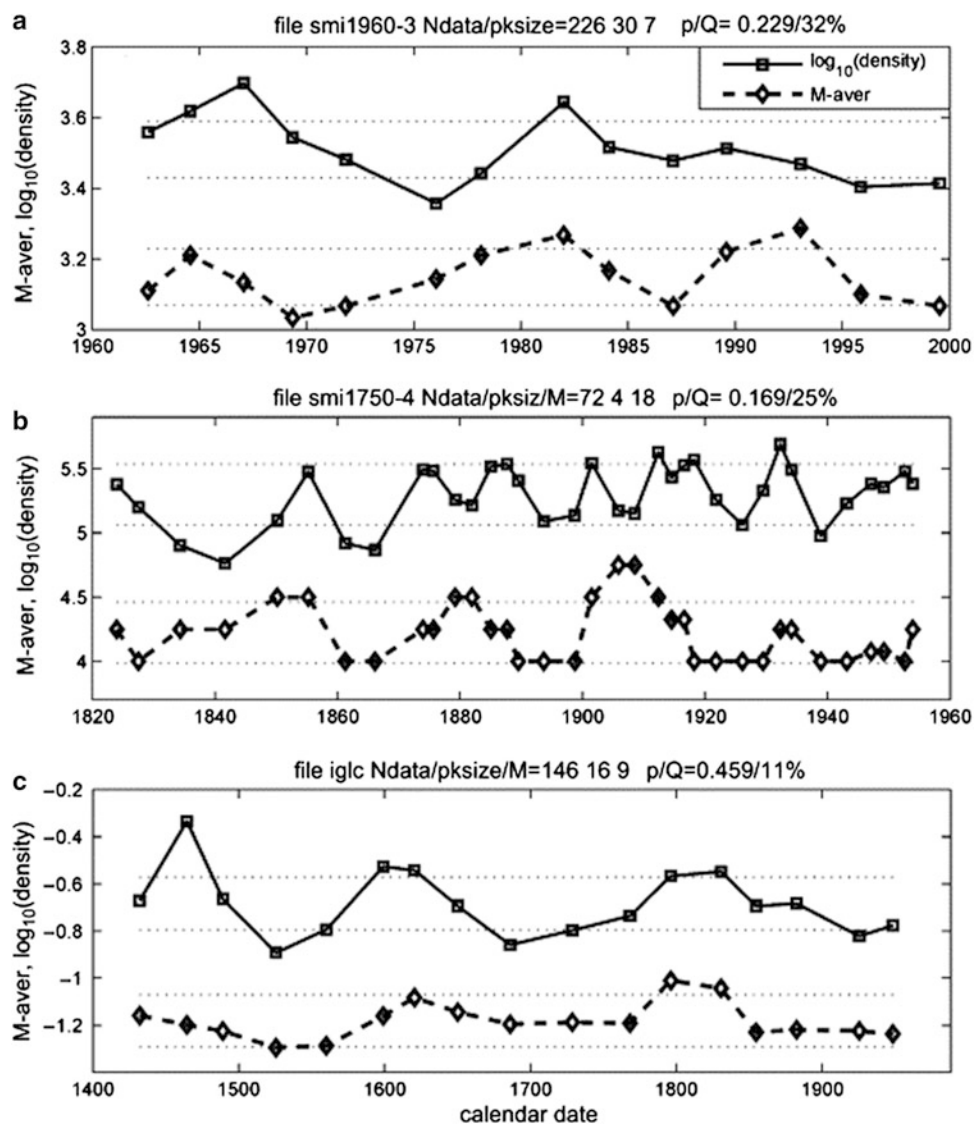
Fractals have basically been used in volcanology as a descriptive technique. They can provide, at the level of a single volcano, some indication of the sort of stochastic model that might fit the observed behavior. Applied to multiple volcanoes, they can provide a similarity index, useful for sorting volcanoes into homogeneous groups.

Volcanic Regimes

Apart from the nonhomogeneous model (4), the parametric models examined above have all been stationary, in the sense that the statistical properties of the intensity function do not vary over time (see also Bebbington 2013; Marzocchi and Bebbington 2012). However, Wadge (1982) postulated that, while a volcano could be in steady state on a timescale of years to decades, in the longer term, the activity can wax and wane. Thus, particularly for long-term hazard forecasts, it is desirable to formulate methods of detecting (and modeling) such nonstationary activity.

Wickman (1966a) suggested that the activity of a volcano could change over time, perhaps in discrete steps, which we will term *regimes*. Mulargia et al. (1987) addressed the importance of objectively identifying regimes of a volcano, noting however that (a) the number of regimes is unknown, (b) they may follow different distributions, and (c) sample sizes are generally small. Different patterns of eruption and magma output are supposedly features of most volcanoes and fundamental to the understanding and modeling of eruptions. These regimes may represent changes in the eruption mechanism, the mechanism for transport of magma to the surface, or the eruptive style (see, e.g., the interpretation by Wadge et al. (1975)). The estimated properties of such regimes can exclude certain models or mechanisms of volcanic eruption (Klein 1982).

Various methods have been used to identify changes in regime, one based on cumulative eruptive volume (Wadge 1982; Wadge et al. 1975) already having been illustrated in Fig. 7. Another



Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 11 Comparison of temporal trends of log event rate (*solid lines and squares*) and of “average magnitude” $M\text{-aver}$ (*dashes and diamonds*), for data sets SMI3 (a), SMI4 (b), and IVI (c), estimated using event packets of constant number. See Table 2 of (Gusev 2008) for details. Plots for $M\text{-aver}$ are shifted vertically for visual clarity.

Horizontal dot lines indicate approximate $\pm 1\sigma$ corridors assuming no temporal variations (Reprinted from Gusev AA (2008) Temporal structure of the global sequence of volcanic eruptions: Order clustering and intermittent discharge rate. Phys Earth Planet Int 166:213 © 2008 Elsevier B.V., with permission from Elsevier)

uses a running mean of time between onsets. Suppose that the individual repose are denoted by $r_i = t_{i+1} - t_i$, $i = 1, \dots, n - 1$, and that $\sum_{i=1}^{n-1} r_i = T$. If the repose are divided into disjoint groups of m consecutive repose, then

$(2(n - 1)/T) \sum_{i=(k-1)m+1}^{km} r_i$, for $k = 1, \dots, \lfloor (n - 1)/m \rfloor$, where $\lfloor \cdot \rfloor$ is the greatest integer function, is independent, and has a chi-square distribution with $2m$ degrees of freedom (Klein 1982).

Mulargia et al. (1987) presented a method based on the theory of change-point problems. Suppose that we have a time series of data $X_1, \dots, X_n$, which might be, for example, inter-onset intervals, eruptive volumes, or effusion rates. The algorithm begins by setting a significance level α and examining the time series to determine the most significant change point. If a change point is found, the time series is divided into two parts at this point, and each examined to determine the most significant change point. The algorithm proceeds recursively until no further significant change points are found. The most significant change point is identified by a search process: For $m = 3, \dots, n - 3$, divide the time series into two segments $X^{(1)} = X_1, \dots, X_m$ and $X^{(2)} = X_{m+1}, \dots, X_n$, and calculate the Kolmogorov-Smirnov (two-tail) statistic as

$$J = \left(\frac{m(n-m)}{n} \right)^{1/2} \max_x |F_1(x; m) - F_2(x; n-m)|, \quad (42)$$

where $F_I(x; k) = \#\{i: X_i^{(I)} \leq x\} / k$ is the empirical distribution function of the first segment, et cetera. For large $m, n - m (> 30)$, the approximate critical value J_α of the statistic can be determined from

$$\Pr(J < J_\alpha) = \sum_{k=-\infty}^{\infty} (-1)^k \exp(-2k^2 J_\alpha^2) = 1 - \alpha. \quad (43)$$

Exact critical values for small $m, n - m$ are also available.

Later approaches used the cumulative count of eruptions in a statistical control chart (Ho 1992b), applied by Burt et al. (1994) to Nyamuragira, and rank order statistics for the size of event (Pyle 1998). Bebbington (2007) noted that a hidden Markov model for the activity, with the unobserved state representing the regime controlling the embedded renewal process, provides for regime identification via the Viterbi algorithm, which finds the most likely path through the hidden states. The mixture of exponentials model (Mendoza-Rosas and De La Cruz-Reyna 2009),

while expressed as a renewal model, is fitted in a manner which is better conceptualized as a regime model in a Bayesian framework.

Mount Etna

Mount Etna possesses one of the most complete records of volcanism known, although the many catalogs differ to some degree and can result in quantitatively different but qualitatively similar analyses (Bebbington 2007; Smethurst et al. 2009; Wang and Bebbington 2013). A variety of eruption styles have been displayed, but since 1500 the activity has been predominately effusive eruptions from lateral vents on the volcano's flanks or from the central craters. The record of summit eruptions is incomplete prior to 1970 (Wadge and Guest 1981), but data for flank eruptions is considered complete since 1600 or so (Mulargia et al. 1987). The volumes of all such eruptions have been estimated (Sandri et al. 2005), and these have been shown to be well correlated with the observed duration (Mulargia et al. 1985). As the summit eruptions appear to be of different style (Wadge et al. 1975) and to have different mechanisms to the flank eruptions (Mulargia et al. 1992), the longer record of the latter can be analyzed in isolation. It has also been demonstrated that the flank eruptions drive the system (Bebbington 2008) and only flank eruptions pose a major threat to inhabited areas (Salvi et al. 2006).

Using the cumulative erupted volume curve, Wadge and Guest (1981) found that the volume output was in steady state 1971–1981. Mulargia et al. (1985) observed that the events 1600–1980 were satisfactorily fit by a stationary Poisson process but that the distribution of eruptive durations changed at the end of the seventeenth century, confirming the observation of Wadge et al. (1975). However, Mulargia et al. (1987) identified a change point in the inter-onset times at 1865, both parts being consistent with a Poisson process, and, when augmented by volume data, change points at 1670 and 1750. Using the effusion rate (=volume/duration), Mulargia et al. (1987) found a change point at 1950 in the 1600–1980 data, with the later sequence being of

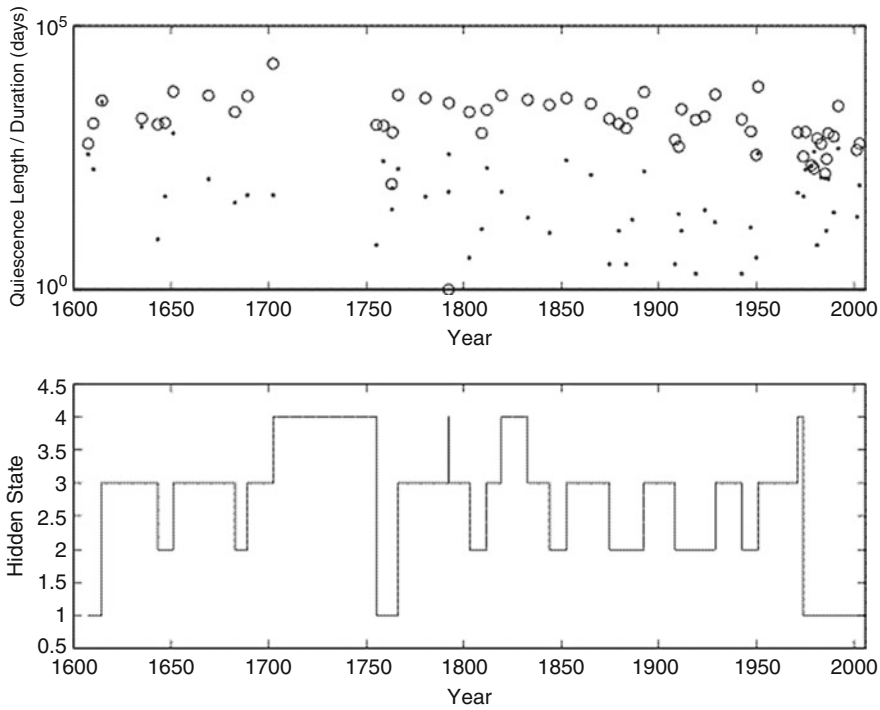
significantly lower rate. A repeat of the analysis by Gasperini et al. (1990) for the data 1978–1987 identified a change point in the inter-onset times at November 1978, although no change point occurred in the volume series. Ho (1992b) fitted a statistical control chart based on the Weibull process and found departures from the trend (Eq. 4) in 1702 and 1759. Marzocchi (1996) found some evidence of a trend in inter-onset times in the 1600–1994 data, while Sandri et al. (2005) found no change points in the inter-onset times or eruptive volumes in the 1971–2002 data. Salvi et al. (2006) determined that the activity 1536–2001 was not Poisson, and although there was no trend 1536–1980, there was a trend in the data 1536–2001. They therefore concluded that the last 20 years of data is from a non-homogeneous Poisson process and fitted a Weibull process model with increasing trend. The idea was put on a more statistical basis by Smethurst et al. (2009) using piecewise linear hazard rates, their fitted hazard having a constant level of 0.11 onsets/year until 1964, after which the fitted hazard rate rises at 0.016/year per year.

The regime change points identified in the Mount Etna flank eruption data are summarized in Table 1. The lack of consistency is due to differences in both models and, via the fitting of them, in the coverage of the data. These models

are “first order” in the data being analyzed. Using a bivariate (repose length/eruption duration) hidden Markov model, Bebbington (2007) was able to examine the “second-order” quantity of the correlation between durations and subsequent repose. The eruption record suggested the existence of four different hidden states or regimes. The most persistent of these corresponded to Poisson behavior, with the others having quasiperiodic repose (long or short eruptions), and finally a very erratic over-dispersed repose length. The conclusion was that changes in volcanic regime may be more frequent and/or fleeting, and thus undetectable by change-point methods, than has been thought. Notably, the identified change points (Fig. 12) corresponded closely to those identified manually by Behncke and Neri (2003), showing that these are statistically identifiable. The complex relation between the duration of an eruption and the subsequent repose appears to be at least partially explained by an open/closed conduit system model with transitions between them. On the other hand, the correlation between repose and subsequent durations may be sensitive to the longer of the two cycles identified by Behncke and Neri (2003), which was also uncovered (Fig. 13) by the volume-history model (Bebbington 2008). This would indicate that flank eruptions are being fed from more than one

Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Table 1 Regime change points for Etna flank eruptions

Data type	Source	Date range	Change points
Onset counts	Mulargia et al. (1985)	1600–1980	None
	Smethurst et al. (2009)	1610–2008	1936 or 1964
Inter-onset times	Mulargia et al. (1987)	1600–1978	1865
	Gasperini et al. (1990)	1978–1987	1987
	Ho (1992b)	1600–1978	1702, 1759
	Marzocchi (1996)	1600–1994	(Trend)
	Sandri et al. (2005)	1971–2002	None
	Salvi et al. (2006)	1536–2001	1980
Volume-repose	Wadge et al. (1975)	1535–1974	1610, 1669, 1759/1763
	Wadge and Guest (1981)	1971–1981	None
	Mulargia et al. (1987)	1600–1978	1670, 1750, 1950
	Gasperini et al. (1990)	1978–1987	None
	Sandri et al. (2005)	1971–2002	None



Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 12 Mount Etna: optimal hidden path for the bivariate Weibull time-predictable HMM. Circles are repose lengths, dots durations (From Bebbington MS

(2007) Identifying volcanic regimes using hidden Markov models. *Geophys J Int* 171: 921–942, Fig. 6, © 2007 The Author, Journal compilation © 2007 RAS)

magma chamber, of widely different sizes and recharge rates.

Spatial Aspects

In the case of polygenetic cones, the major aspect of hazard modeling is temporal. Spatial aspects are limited to the direction of flank eruptions (cf. Salvi et al. 2006) and the likely paths of lava flows and lahars. In monogenetic volcanic fields, where cones generally correspond to single eruptions, there is a true spatiotemporal element to modeling occurrence data.

Spatiotemporal Intensities

The models used to describe spatiotemporal eruption occurrence have been largely nonparametric or kernel type. Connor and Hill (1995) gave three methods:

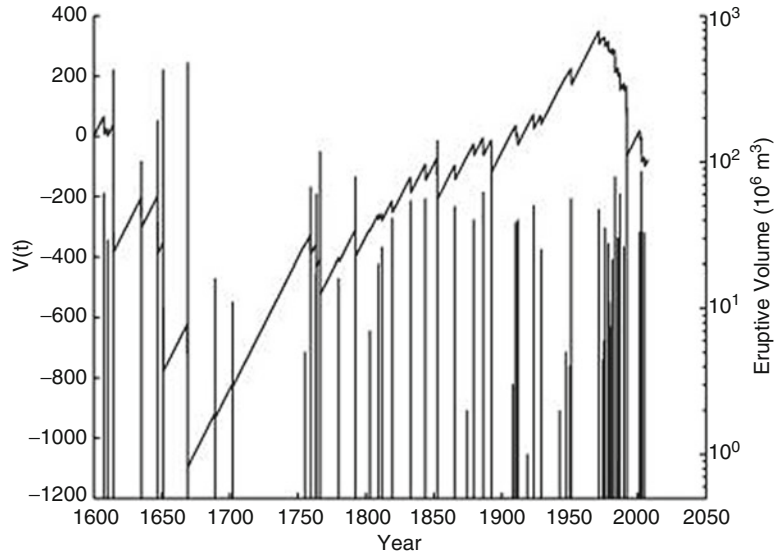
1. *Spatiotemporal nearest-neighbor estimate:* Suppose we have n volcanoes, that the formation of the i th volcano occurred at time t_i , and let u_i be the area of a circle with radius equal to the distance between the point x of interest and the i th volcano. Further, let $j = 1, \dots, m$ index the m th nearest neighbors to the point x using the distance metric $u_i(t - t_i)$. Then the estimated intensity at the point x is

$$\lambda(x, t; m) = \frac{m}{\sum_{j=1}^m u_j(t - t_j)}, \quad (44)$$

where m is the number of nearest neighbors used in the estimation, which was set equal to the number of volcanoes, although a smaller choice is possible. The temporal hazard for an area A can then be obtained by integrating the intensity (Eq. 44) over the area. Note that in

Volcanic Eruptions: Stochastic Models of Occurrence Patterns,

Fig. 13 Eruptive volumes (vertical bars) and fitted stored volume (curve) for flank eruptions of Mt Etna (Reprinted from Bebbington M (2008) Incorporating the eruptive history in a stochastic model for volcanic eruptions. *J Volcanol Geotherm Res* 175:328, © 2008 Elsevier B.V., with permission from Elsevier)



practice, to avoid singularities, $u_i > c$, a constant, and the intensity is summed over a grid. The choice of c can be an interesting question. Condit and Connor (1996) proposed that the optimal m can be determined by matching the average recurrence rate $\int_{x \in A} \lambda(x, t; m) dx$ to the observed average recurrence rate $\bar{\lambda}(t) = N(t - s, t) / As$, in a window of length s , for the area A . Further, Condit and Connor (1996) also suggested that overestimation during rapidly waning stages of activity can be addressed by defining a threshold time beyond which prior eruptions are not included in the calculation.

2. *Kernel estimate*: If we have n volcanoes, then using the Epanechnikov kernel,

$$\kappa_i(h) = \max \left\{ 0, \frac{2}{\pi} \left[1 - \left(\frac{d_i}{h} \right)^2 \right] \right\}, \quad (45)$$

where d_i is the distance from x to the i th volcano, we have the spatial density (or PDF),

$$\lambda(x) = \frac{1}{e_h} \sum_{i=1}^n \frac{\kappa_i(h)}{n}, \quad (46)$$

where e_h is an edge correction. Conway et al. (1998) used the Gaussian kernel

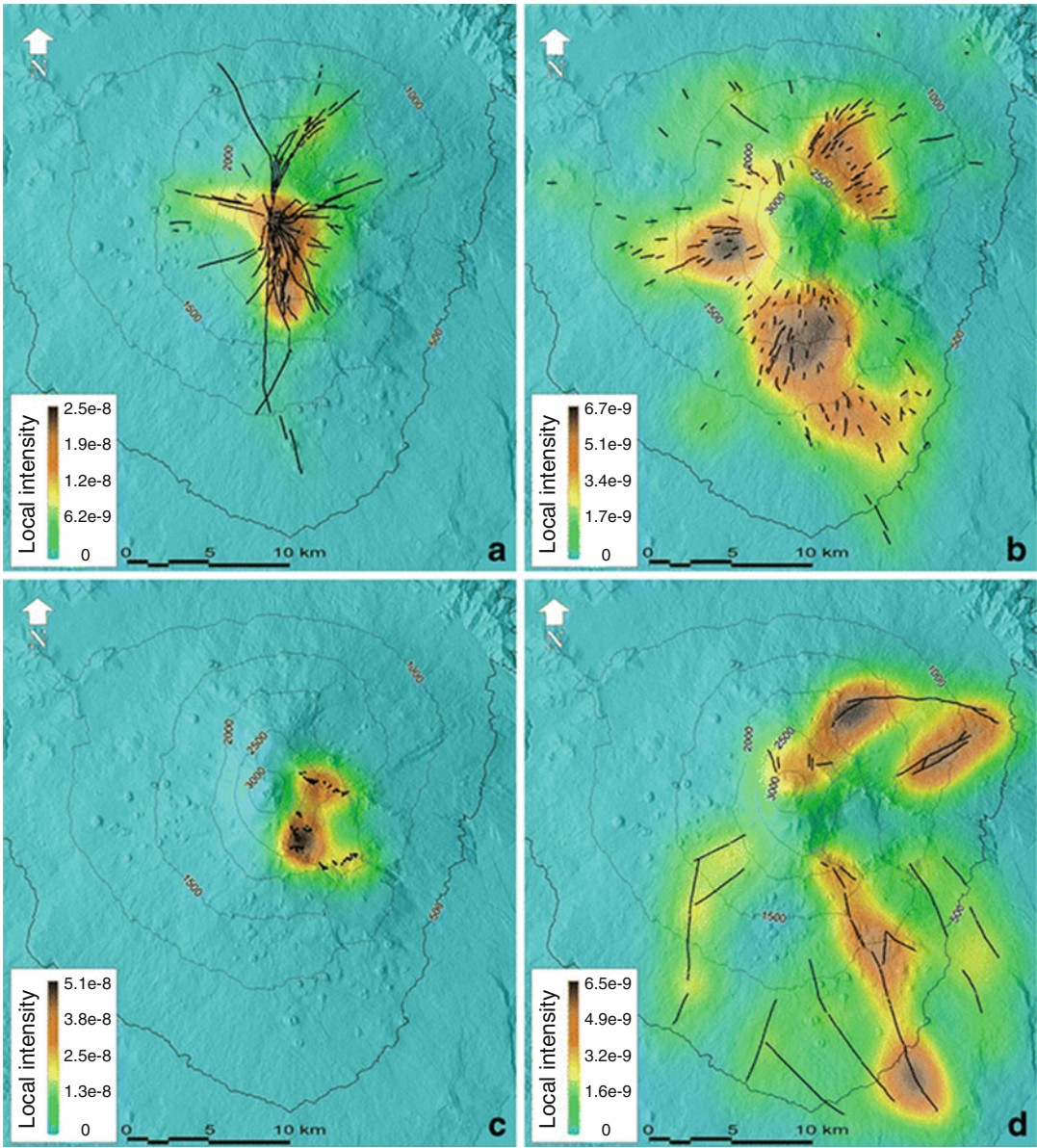
$$\kappa_i(h) = \frac{1}{2\pi h^2} \exp \left[-0.5(d_i/h)^2 \right] \quad (47)$$

instead. The parameter h is a smoothing constant. A small value of h concentrates the probability close to existing volcanoes, while a large value distributes it more uniformly. Estimating the best value of h is a difficult task – see Vere-Jones (1992) for an earth science-based discussion, but cross validation (Duong 2007) is often used. The procedure has been automated in a GIS framework allowing incorporation of point and other data by Felpeto et al. (2007) and used to estimate the hazard on Tenerife (Marti and Felpeto 2010). A similar exercise has been conducted for the susceptibility of flank vents at Mt Etna (Fig. 14) by Cappello et al. (2012).

The kernel in Eq. 46 is isotropic, i.e., radially symmetric. An anisotropic kernel density estimate

$$\eta(x) = \frac{1}{n2\pi\sqrt{|H|}} \sum_{i=1}^n \exp \left[-0.5(x - x_i)^T H^{-1}(x - x_i) \right], \quad (48)$$

where H is a 2×2 matrix, has been suggested (Connor and Connor 2009) and implemented by Kiyosugi et al. (2010) for the Abu Monogenetic Volcanic Group and by Bebbington



Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 14 PDFs for post-1600 fractures (a), pre-1600 fractures (b), dikes (c), and faults (d) calculated with a Gaussian kernel (With kind permission from Springer

Science + Business Media: Bull Volcanol, Spatial vent opening probability map of Etna volcano (Sicily, Italy), vol 74, 2012, p 2090, Cappello A et al., Fig. 6, © Springer-Verlag 2012)

and Cronin (2011) for the Auckland Volcanic Field. Again H can be estimated by Least Squares Cross-Validation (LSCV) or by the Sum of Asymptotic Mean Squared Error (SAMSE) (Duong and Hazelton 2003). The

latter is more stable, but the trade-off is that it tends to obscure possible fine-scale spatial structure.

3. *Nearest-neighbor kernel estimate*: Here we replace the smoothing constant h in the Kernel

estimate with $d_{(m)}$, the distance to the m th nearest neighbor using the d_i metric.

The two (spatial) kernel estimates above can then be made into spatiotemporal estimates by multiplying them by a spatially independent temporal rate $\hat{\lambda}_0 = N/T$, resulting in

$$\lambda(x, t) = \lambda(x) \hat{\lambda}_0. \quad (49)$$

Ho and Smith (1998) generalized (Eq. 49) by replacing $\hat{\lambda}_0$ by the Weibull process intensity (Eq. 4). A method of evaluating spatiotemporal forecasts from these and other, notably a uniform spatial element, has been proposed (Bebbington 2013) and illustrated on data (Bebbington and Cronin 2011) from the Auckland Volcanic Field.

Martin et al. (2004) extended the spatiotemporal idea with a Bayesian formulation to incorporate information from other geophysical data such as P-wave velocity perturbations or geothermal gradients. The spatiotemporal intensity (Eq. 46) is used as an a priori intensity $P(x)$, and a likelihood function $L(\theta|x)$ is generated by conditioning the geophysical data θ on the locations of volcanic events. The a posteriori intensity can then be obtained from Bayes' theorem as

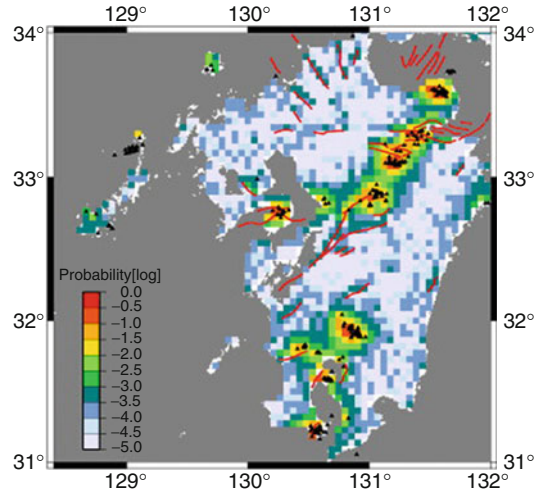
$$P(x|\theta) = \frac{P(x)L(\theta|x)}{\int_{y \in A} P(y)L(\theta|y) dA}, \quad (50)$$

where A is the volcanic field. An example using gravity data can be found in Weller (2004).

The Bayesian event tree formulation has also been used (Selva et al. 2012) to estimate spatial probabilities, but here the prior is constructed from the geophysical data (the caldera floor, tectonic structures, and eruptive vent structures) and is then modified by the observed record. At a similar scale, Jaquet et al. (2012) have extended the Cox process to modeling the spatial distribution of volcanism (Fig. 15).

Markov Chains

In the case where there is a geologically useful spatial classification, and sufficient data, a simple

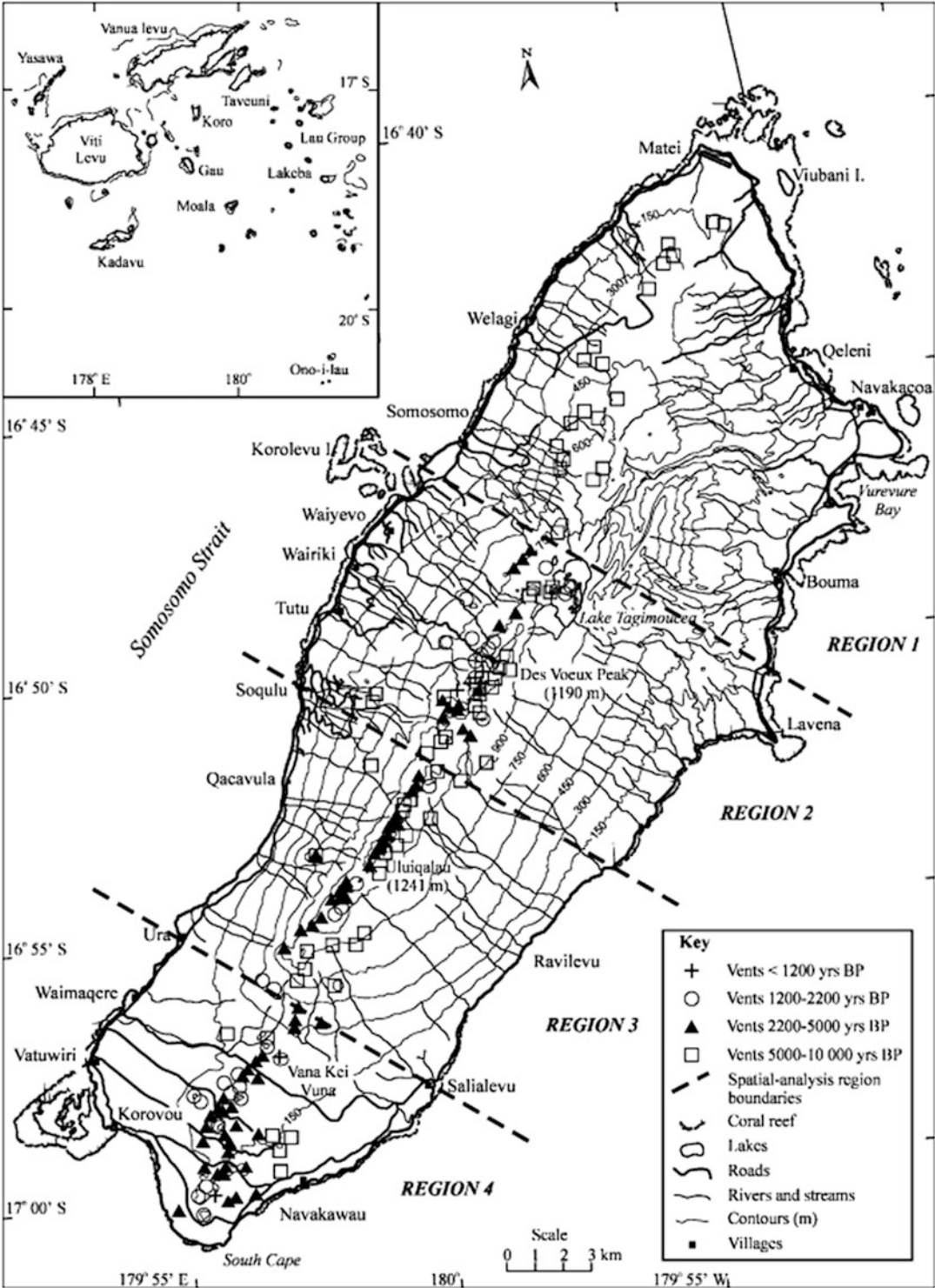


Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 15 Volcanic hazard map for Kyushu displaying the probability of one or more volcanic events for the next 100,000 years using a 5×5 km domain (case I) (Reprinted from Jaquet O, Lantuejoul C, Goto J (2012) Probabilistic estimation of long-term volcanic hazard with assimilation of geophysics and tectonic data. *J Volcanol Geotherm Res* 235–236:34 © 2012 Elsevier B.V., with permission from Elsevier)

Markov chain approach can be used. If the i th eruption occurs in spatial region X_i , then the transition matrix is $P = (p_{jk})$, where $p_{jk} = \Pr(X_{i+1} = k | X_i = j)$. The maximum likelihood estimates are $\hat{p}_{jk} = N_{jk} / \sum_l N_{jl}$, where N_{jk} is the number of transitions from j to k observed in the data. The Markov chain can be tested against a null hypothesis of independence using the statistic

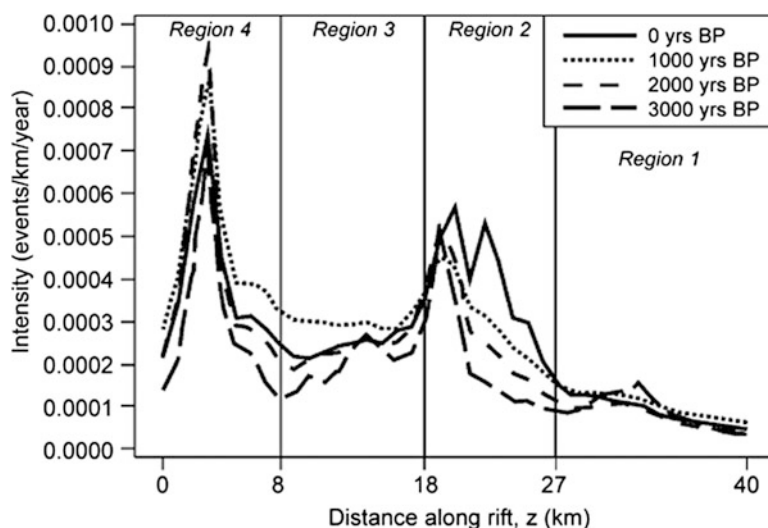
$$\chi^2 = 2 \sum_j \sum_k N_{jk} \log \left[\frac{N_{jk} / \sum_l N_{jl}}{\sum_l N_{lk} / \sum_l \sum_m N_{lm}} \right], \quad (51)$$

which asymptotically has an χ^2 distribution with $(K-1)^2$ degrees of freedom, where K is the number of states. Cronin et al. (2001) used this approach, made feasible by the essentially linear rift formation shown in Fig. 16, to estimate the hazard at the heavily populated southwest end of Taveuni, Fiji. This was combined (cf. Eq. 49) with a hierarchical renewal model for the temporal aspect to provide



Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 16 Location map of Taveuni (inset Fiji group), with main villages, positions of Holocene mapped vents, and regions used for the spatial analysis (From

Cronin S, Bebbington M, Lai CD (2001) A probabilistic assessment of eruption recurrence on Taveuni volcano, Fiji. *Bull Volcanol* 63:275, Fig. 1, © Springer-Verlag 2001. With kind permission of Springer Science and Business Media)



Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 17 Plot of spatiotemporal intensities (events/km year⁻¹) over the length of the Taveuni rift zone and at different times between 3000 B.P. and the present (From Cronin S, Bebbington M, Lai CD

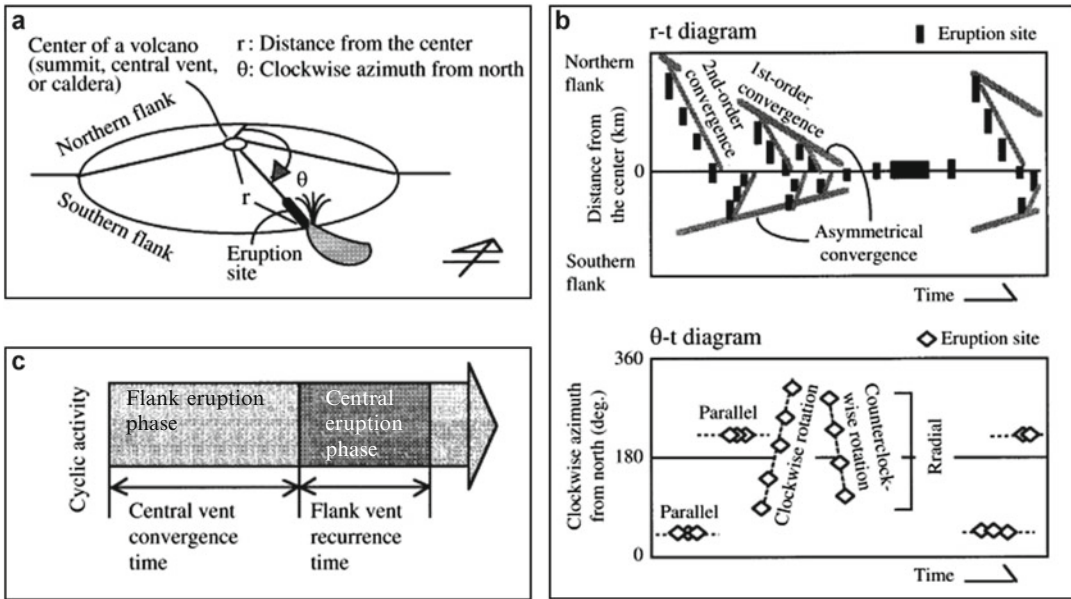
(2001) A probabilistic assessment of eruption recurrence on Taveuni volcano, Fiji. *Bull Volcanol* 63:285, Fig. 8, © Springer-Verlag 2001. With kind permission of Springer Science and Business Media)

an alternative to the spatiotemporal hazard estimate from Eq. 44 shown in Fig. 17. Eliasson et al. (2006) likewise divided the Katla caldera, Iceland, into three sectors to examine the spatial progression of volcanogenic flood events.

In the same vein, a few studies have considered possible correlations between multiple vents on the same volcano. Klein (1982) showed using a runs test that eruptions and intrusions at Kilauea tend to cluster in time and that intrusions occur in place of eruptions during long repose. By dividing Kilauea repose into summit-summit, summit-flank, flank-summit, and flank-flank pairs, and finding that their distributions did not differ statistically, Klein (1982) showed that the location of the previous event provided no information about the time of the next eruption. At Mauna Loa, however, repose following flank eruptions were significantly longer than those following summit eruptions, consistent with the fact that the former are generally more voluminous than the latter. Examination of the sequences of summit and flank eruptions disclosed that Kilauea summit eruptions tend to occur in runs, due solely to the long summit sequence of 1924–1954, while Mauna Loa displays no tendency for clustering

or alternation of summit and flank eruptions. However, Lockwood (1995), in an analysis of 170 well-dated prehistoric lava flows from Mauna Loa, identifies cycling between summit overflows and flank eruptions with a periodicity of about 2,000 years.

Takada (1997) proposed using time-series plots, as shown in Fig. 18, to visually express the temporal relationship between flank and summit eruptions, in particular plots of flank-vent distance and direction from the central vent. Central eruption phases, defined as periods where at least 70 % of eruptions are central vent eruptions without accompanying flank eruptions, alternate with flank eruption phases. Salvi et al. (2006) considered the azimuth distribution of flank eruptions from Mount Etna, as in Fig. 19, determining that there is no significant difference in the pre- and post-1536 distributions. Cappello et al. (2012) made this invariance assumption for post-1600 vents in estimating a spatial probability distribution via kernel smoothing for the site of the next flank eruption, although Smethurst et al. (2009) showed that vents became more closely focused (around the Southeast Crater) following its formation in 1971.



Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 18 (a) Parameters representing the position of an eruption site. The flank of a volcano is divided into two regions, e.g., the northern flank and the southern one. (b) Two types of time-series diagram for eruption sites, representing the distance of eruption sites from the center (r - t diagram) and the clockwise azimuth from the

north (θ - t diagram), respectively. (c) Schematic cyclic activity consisting of flank eruption and central eruption phases (With kind permission from Springer Science+Business Media: Bull Volcanol, Cyclic flank-vent and central-vent eruption patterns, vol 58, 1997, p 540, Takada A, Fig. 1, © Springer-Verlag 1997)

Klein (1982) also examined the relation between activity at the nearby volcanoes of Kilauea and that of Mauna Loa. Separating the record into three parts at the longest repose of each volcano disclosed that the activities were inversely related. The rank correlation of the number of events in 5-year periods 1915–1980 was -0.646 , significant at a level of 0.01, which indicates that they share the same magma source. Bebbington and Lai (1996b) repeated this analysis for the New Zealand volcanoes Ruapehu and Ngauruhoe, finding that the combined eruption sequence was indistinguishable from a superposition of independent Poisson processes. This independence of the eruption sequences argues for separate magma sources.

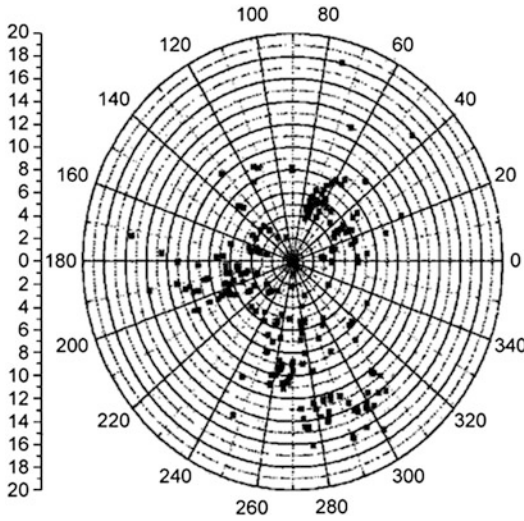
Alignments and Clusters

There has also been considerable interest in detecting alignments or clusters of volcanic

vents, in the hope that these can be related to geological features.

Lutz (1986) proposed the “two-point azimuth method,” where the azimuth between every two vents is measured, generating $n(n-1)/2$ measurements for n vents. These are then binned in 10° intervals, although the bin size could vary, and the result tested for departure from randomness. As noncircular fields and/or a nonhomogeneous density of points can produce a preferred orientation, a correction must be made in these cases. One way of doing this is to use Monte Carlo simulation of random points to produce a reference distribution from which departures can be detected. Lutz and Gutmann (1995) noted that using kernel smoothing methods for the vent location provides an improved distribution from which to perform the Monte Carlo simulation.

Wadge and Cross (1988) suggested the use of the Hough transform. In this technique, a point (x, y) is converted into the normal parameterization (ρ, θ) , where $\rho = x \cos \theta + y \sin \theta$. Thus, each point



Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 19 Azimuthal distribution of cones of lateral eruption on Etna. The center is located on “La Voragine” central crater. The scale on the left is in km from the center (Reprinted from Salvi F, Scandone R, Palma C (2006) Statistical analysis of the historical activity of Mount Etna, aimed at the evaluation of volcanic hazard. J Volcanol Geotherm Res 154:160, © 2006 Elsevier B.V., with permission from Elsevier)

generates a curve in the (ρ, θ) plane as θ varies, and multiple (more than two curves) intersections of these define the alignment of colinear points. In practice the scales for ρ and θ are quantized into discrete bins, which allows for a small degree of departure from strict colinearity. Significance levels can be obtained via Monte Carlo techniques.

Connor (1990) suggested the use of uniform kernel density fusion cluster analysis, using the distance metric

$$d(i, j) = \begin{cases} \frac{1}{2} \left(\frac{1}{m_i(w)} + \frac{1}{m_j(w)} \right), & \|x_i - x_j\| < W \\ \infty, & \text{otherwise,} \end{cases} \quad (52)$$

where $\|\cdot\|$ denotes map distance and $m_i(w)$ is the number of cones within a circle of radius w centered on the i th vent. Cones are linked into clusters starting with the smallest values of d . Clusters are only combined if the “density fusion”

d between the clusters is greater than the maximum d between any two events within either cluster. This makes it possible to recognize overlapping clusters. The search radius w is determined experimentally.

Magill et al. (2005) measured vent clustering in the Auckland Volcanic Field using the statistic

$$L(d) = \sqrt{\frac{K(d)}{\pi}} - d, \quad (53)$$

where

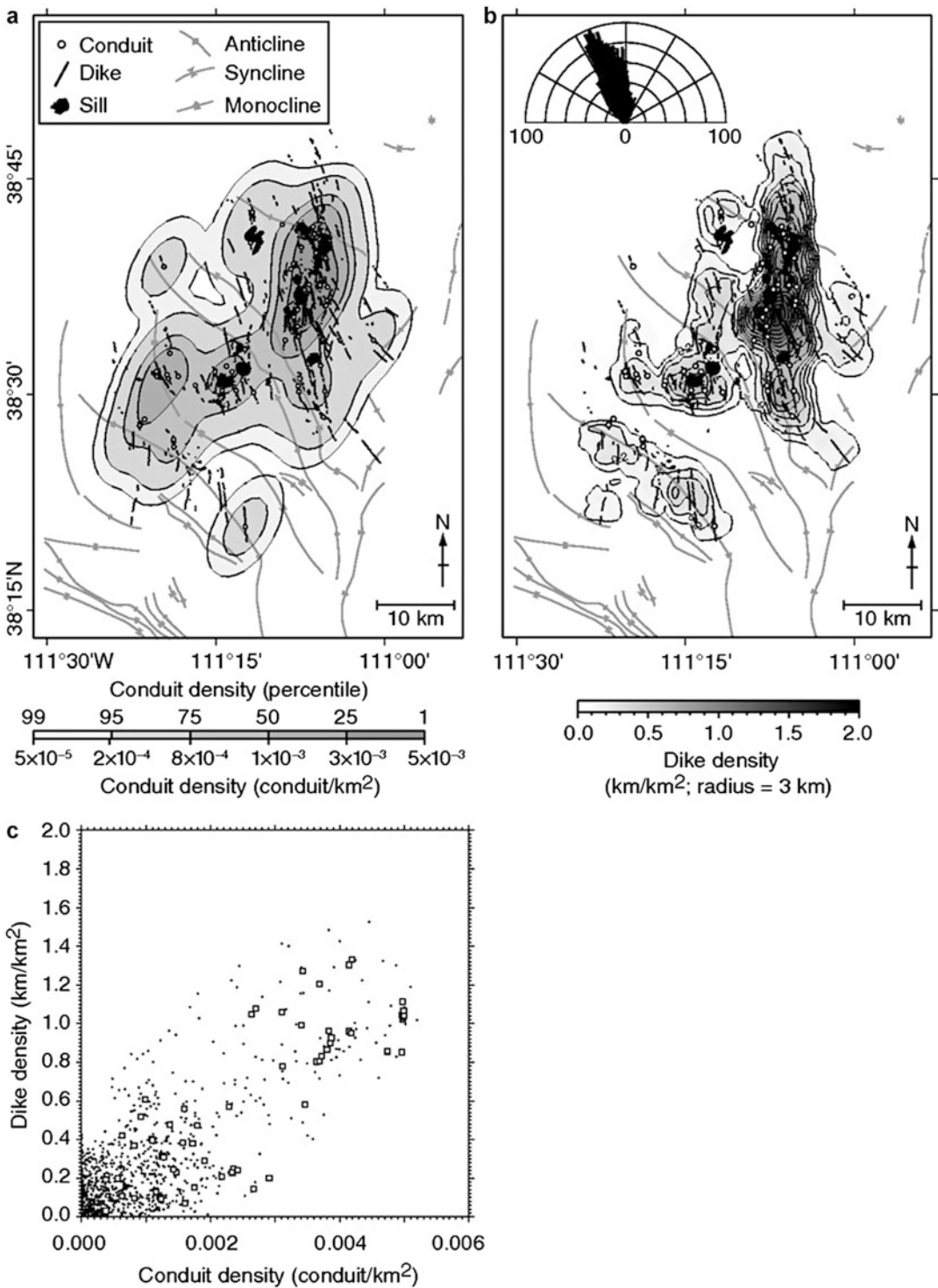
$$K(d) = \frac{A}{n} \sum_{j=1}^n \sum_{i \neq j} \frac{H(d - d_{ij})}{w_{ij}} \quad (54)$$

is Ripley’s K function, A is the area of the field, and d_{ij} is the distance between the i th and j th vents, $H(\cdot)$ is the Heaviside function, and w_{ij} is an edge correction. Peaks of positive $L(d)$ indicate clustering, and troughs of negative values regularity, at that distance. The Auckland field exhibits clustering at between 900 and 1,600 m and spatial regularity at about 4,600 m. This clustering was used to collapse 49 cones into 18 events, but this analysis was compromised by the circular reasoning built into the data, where closeness in space was used to infer closeness in time. A later analysis (Bebbington and Cronin 2011), using a Monte Carlo approach and additional data sources, found no evidence of spatiotemporal dependence.

Wadge and Burt (2011) updated the analysis of Burt et al. (1994) and correlated fissure directions with seismic and tectonic data, while Kiyosugi et al. (2012) have examined the spatial interrelationship between dikes and conduits (Fig. 20), concluding that dikes cluster near conduits with >99 % confidence.

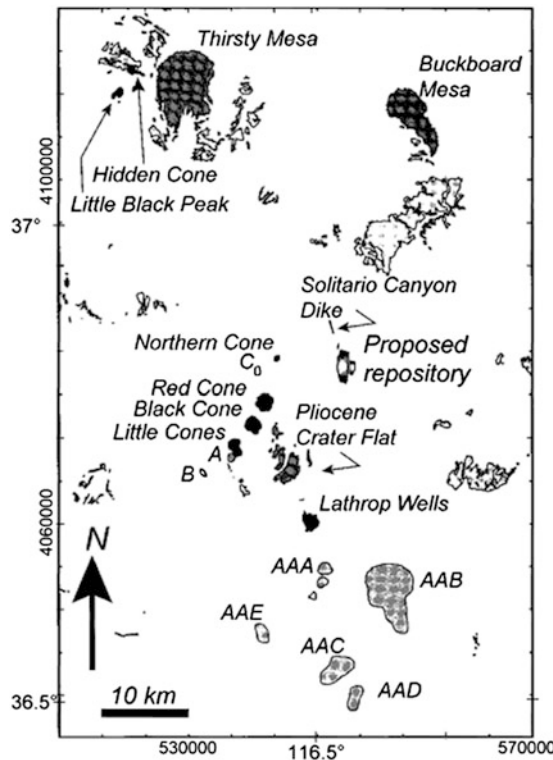
Yucca Mountain

After Mount Etna, the next most studied volcano is probably the volcanic field at Yucca Mountain, Nevada (cf. Fig. 21), due to the high-level



Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 20 Conduit and dike density. (a) Conduit density map. (b) Dike density map. Rose diagram shows

orientation of dike segments. (c) Comparison of conduit and dike densities. Dots show densities at grid points for the main map area. Open squares show densities at conduit



Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 21 Distribution of basaltic rocks in the Yucca Mountain region: *light gray shading*, Miocene; *dark gray shading*, Pliocene; *black shading*, Quaternary. Magnetic anomalies are shown in *light gray shading*, but only AAB is dated (Pliocene). Location is given in latitude/longitude and in Universal Transverse Mercator (UTM)

coordinate pairs (in meters) (From Connor CB, Stamatakis JA, Ferrill DA, Hill BE, Ofoegbu GI, Conway FM, Sagar B, Trapp J (2000) Geologic factors controlling patterns of small-volume basaltic volcanism: Application to a volcanic hazards assessment at Yucca Mountain, Nevada. *J Geophys Res* 105:418 Fig. 1b, © 2000 by the American Geophysical Union)

radioactive waste repository planned there. The available data (Coleman et al. 2004) consists of a large number of basaltic volcanoes, of ages 80,000–10,500,000 years. These of course have large dating errors associated with them, but most analyses do not seem to have included this. The object is to estimate the probability of disruption of the repository by a volcanic intrusion. This is dependent on the probability of an intrusion, the distribution of the intrusion's dimensions, and dimensions of the repository itself.

The first analysis (Crowe et al. 1982) used a Poisson process for the probability of an event, coupled with a Bernoulli probability of an event disrupting the repository:

$$\begin{aligned} \text{Pr}(\text{no disruption before time } t) &= \sum_{n=0}^{\infty} \frac{(\lambda t)^n}{n!} (1-p)^n \\ &= e^{-\lambda t p}. \end{aligned} \tag{55}$$

Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 20 (continued) locations (From Kiyosugi K, Connor CB, Wetmore PH, Ferwerda BP, Germa AM, Connor LJ, Hintz AR (2012) Relationship

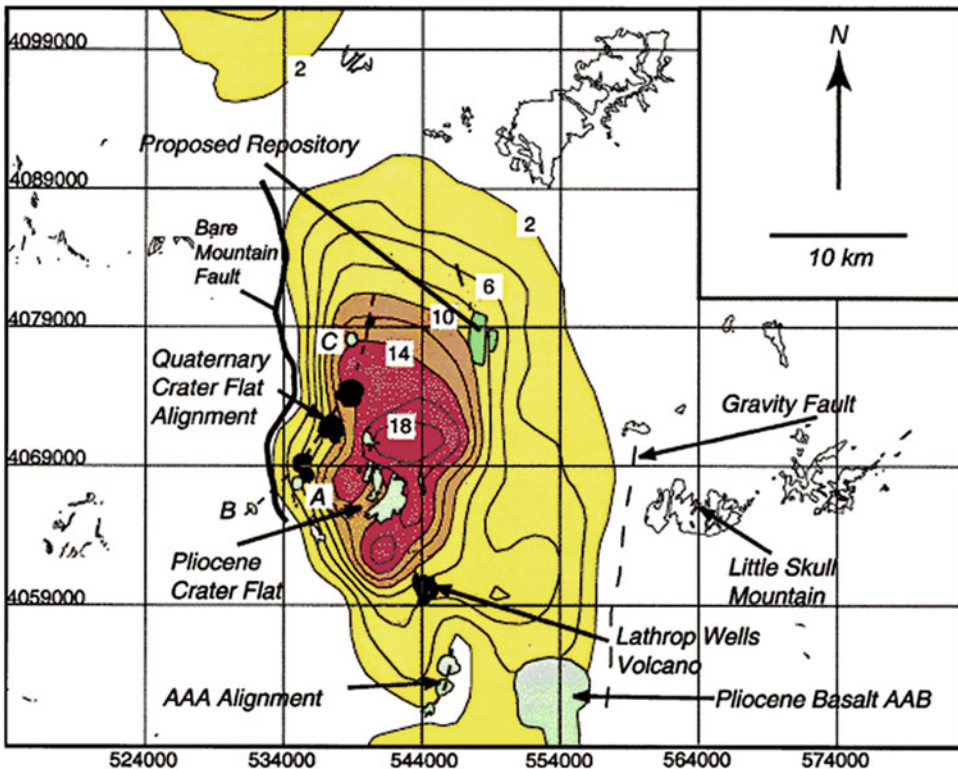
between dike and volcanic conduit distribution in a highly eroded monogenetic volcanic field: San Rafael, Utah, USA. *Geology* 40:695–698, Fig. 2)

The probability of an event being disruptive was estimated as $p = a/A$, where a is the area of the repository and A the area, including the repository, used to define λ . The latter was estimated as N/T , where N is the number of scoria cones and T the corresponding time period. Ho et al. (1991) elaborated on means of estimating λ through event counts, repose intervals, and magma volumes, and Ho (1992a) suggested a Bayesian framework for estimating p .

The next approach (Ho 1992a) attempted to incorporate a trend in the rate of volcanism using the Weibull process (Eq. 4). As the data consisted of a number of clusters each containing several cones, but only the former had been dated, the

event dates input to the Weibull process consisted of multiples of each of a few dates. Depending on the details of the implementation, there was either a strong ($\beta > 2$) increasing trend or no significant trend at all. Notably, Ho (1992a) used a homogeneous Poisson process, with rate equal to the fitted Weibull process intensity at the present, for future prediction, although a continuing Weibull process was one possibility considered in the sensitivity analysis (Ho 1995). Ho and Smith (1997) combined the Weibull process and Bayesian estimation of p in a formulation allowing the incorporation of “expert knowledge.”

Sheridan (1992) included a spatial element by fitting a bivariate Gaussian distribution with five

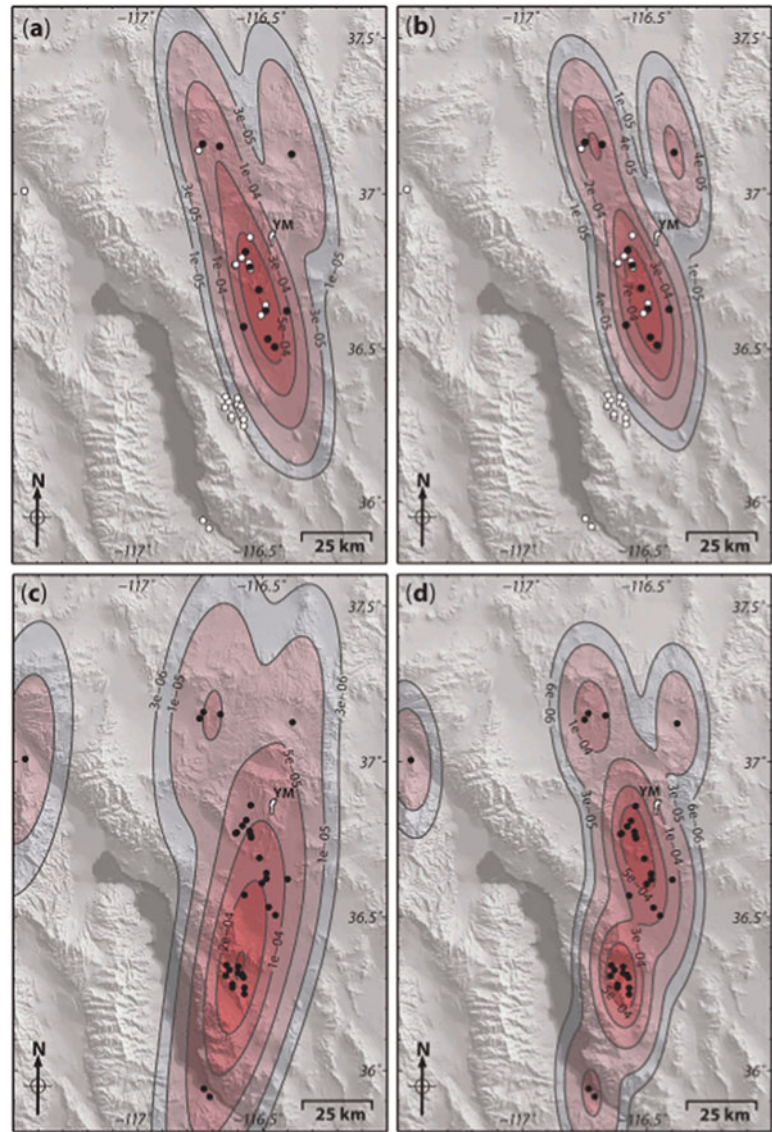


Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 22 The spatial recurrence rate (volcanic events/km²) contoured for the Yucca Mountain region, based on the distribution of Quaternary volcanism and its relationship to the Bare Mountain Fault. The contour interval is 2×10^{-4} volcanic events/km². Location is given in Universal Transverse Mercator (UTM) coordinate pairs

(in meters) (From Connor CB, Stamatakis JA, Ferrill DA, Hill BE, Ofoegbu GI, Conway FM, Sagar B, Trapp J (2000) Geologic factors controlling patterns of small-volume basaltic volcanism: Application to a volcanic hazards assessment at Yucca Mountain, Nevada. J Geophys Res 105:422 Plate 2, © 2000 by the American Geophysical Union)

**Volcanic Eruptions:
Stochastic Models of
Occurrence Patterns,**

Fig. 23 Spatial density estimates of potential volcanism about Yucca Mountain (YMR event data set using (a) SCV and (b) SAMSE methods; AVIP data set using (c) SCV and (d) SAMSE methods). Contours at 25th, 50th, 75th, 95th, and 99th percentiles (e.g., on map (a) locations within the $1e-04$ contour (i.e., 75th quartile) have spatial density $>1 \times 10^{-4} \text{ km}^{-2}$; given a volcanic event, there is a 75 % chance it will occur within this quartile) (Reprinted from Connor CB, Connor LJ (2009) Estimating spatial density with kernel methods. In Connor CB, Chapman NA, Connor LJ (Eds) Volcanic and tectonic hazard assessment for nuclear facilities. Cambridge University Press, Cambridge, pp 346–368, plate 14.3. © Cambridge University Press 2009)



parameters (the (x,y) coordinates of the center, the length of the major and minor axes, and their orientation) to the volcanic field as a whole. Connor and Hill (1995) used the spatiotemporal estimate (Eq. 44) and integrated the resulting intensity over the area of the proposed repository, using a gamma distribution for the area of a new volcano. The effects on the intensity of future events beyond the vicinity of the repository were ignored. Similar results were obtained using

kernel-type estimators. Ho and Smith (1998) elaborated on the latter by making the independent temporal component a Weibull process, while Connor et al. (2000), while retaining the temporal Poisson process component, imposed a variety of geological constraints from crustal density, vent and fault alignments, and vent positions, on the spatial Gaussian kernel intensity in Fig. 22. A sensitivity analysis restricting the vent locations to various date subsets was also conducted. A later

Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Table 2 Summary of hazard assessment for Yucca Mountain

Source	Estimation method		Probability of disruption in 10^5 years
	Temporal	Spatial	
Crowe et al. (1982)	Poisson process	$p = a/A$	10^{-5} to 10^{-3}
Ho et al. (1991)	Poisson process	(p)	$0.5p$
Sheridan (1992)	(λ)	Bivariate Gaussian	$6\lambda \times 10^{-3}$ to $1.7\lambda \times 10^{-2}$
Ho (1992a)	Weibull (past)/Poisson (future) process	Bayesian	10^{-3} to 7×10^{-3} (10^4 years)
Ho (1995)	Weibull process	Bayesian	2×10^{-5} to 7×10^{-3}
Connor and Hill (1995)	Spatiotemporal		10^{-4} to 5×10^{-4} (10^4 years)
Connor and Hill (1995)	Poisson process	Kernel	10^{-4} to 5×10^{-4} (10^4 years)
Geomatrix (1996)	Expert panel: various		5×10^{-5} to 5×10^{-3}
Ho and Smith (1997)	Weibull process	Bayesian	10^{-3} to 3×10^2
Crowe et al. (1998)	Summary: various		7×10^{-5} to 4×10^{-3}
Ho and Smith (1998)	Weibull process	Kernel	10^{-4} to 6×10^{-4}
Connor et al. (2000)	Weighted spatiotemporal		10^{-3} to 10^{-2}
Coleman et al. (2004)	Poisson process		10^{-3} to 10^{-2}

analysis (Connor and Connor 2009) looked at the effects of both anisotropic bandwidth selection (and the estimator used) and data selection (Fig. 23), concluding that the inclusion/exclusion of Deathwater and Greenwater events affected the spatial hazard more than the bandwidth used, even without taking into account the effects on the temporal recurrence rate.

Crowe et al. (1998), in their “final report” of hazard studies at Yucca Mountain, summarized previous work, in particular that of Geomatrix Consultants (1996). They also fitted uniform, triangular (two varieties), and normal distributions to event count data and simulated the results to obtain hazard estimates.

Smith et al. (2002) suggested that the volcanism in the Yucca Mountain area could be episodic and linked to an area of hot mantle. This would make possible another future peak of volcanism, with consequent underestimation of the recurrence rate. It was suggested that recurrence rates up to four times those previously used might be feasible. On the other hand, Coleman et al. (2004) considered the lack of any detected dikes in or above the potential

repository block, treating intrusions as an isolated Poisson process with zero observations in 13,000,000 years (although there was one “near miss”). The conclusion was that high probabilities (10^{-6} per year) of disruption are “unrealistic”; Table 2 summarizes the published hazard estimates.

Interactions with Earthquakes

A common question (Bebbington and Marzocchi 2011; Manga and Brodsky 2006) is whether an eruption is related to a preceding large regional earthquake (or vice versa). Stress changes from earthquakes resolved at dike locations tend to be small, in fact smaller than those associated with solid earth tides. Those earthquakes within a few fault lengths are an exception. However, small increments of stress may of course advance the occurrence of an already imminent eruption. On a longer scale, as both eruptions and earthquakes are tectonic in nature, they must be related. Any such correlation may provide additional information about the likelihood of an imminent eruption

which may be caused, or at least “detected,” by seismic activity. Alternatively, one might be interested in the likelihood of potentially dangerous earthquakes during a period of volcanic activity.

Sharp et al. (1981) considered the relationship between eruptions of Mount Etna and earthquakes from various structural zones of southern Italy. Both the earthquakes and eruption onsets were treated as Poisson processes, an assumption that had to be tested, with intensities λ_a and λ_b , respectively. A variable $\xi(t)$ is formed by summing the number of A series events in the intervals $(B_i, B_i + t)$, where (B_i) are the observed times of the B-series events. If the two series are independent, then $\xi(t)$ should have a Poisson distribution with mean $\hat{\lambda} = \hat{\lambda}_a \hat{\lambda}_b T$, where T is the length of the observation period, and $\hat{\lambda}_i = N_i(T)/T$, for $i = a, b$. For example, if B-series events consistently lead to A-series events by some interval, less than t , $\hat{\xi}$ will be significantly larger than $\hat{\lambda}$. The significance of an observed $\hat{\xi}(t)$, or confidence limits, can be obtained from the null distribution, which is Poisson with mean $\hat{\lambda}$. Repeating the procedure for varying t identifies significant correlations. Sharp et al. (1981) found that the earthquakes were not Poisson, but could be represented as a nonstationary Poisson process. A significant relationship was found between local earthquakes and subsequent flank eruptions within a few hundred days. Also, summit eruptions were found to occur before flank eruptions not preceded by earthquakes, but the summit eruptions themselves were not correlated with earthquakes. This suggests two mechanisms for flank eruptions of Etna: Those preceded by summit eruptions may be caused by magma pressure in the reservoir, while those preceded by earthquakes may be due to fracturing of the flank by earthquakes caused by tensile forces associated with the east-west extension of eastern Sicily. Feuillet et al. (2006) isolated normal faults and showed that there was a two-way coupling, with correlation between the magnitude of the earthquake and the rate of lava eruption.

Klein (1984) examined various precursors of volcanic eruptions at Kilauea by examining the ratio of these precursors preceding an eruption to

their long-term average value. The candidates included summit tilt, daily counts of small caldera earthquakes, repose since the last eruption, earth tides, and rainfall, as shown in Fig. 24. The number of caldera earthquakes was highly significant as a short-term (10 days or less) predictor, but tilt rate was a better predictor beyond 20 days. The fortnightly earth tidal modulation was also significant (cf. Martin and Rose 1981), as it appears to trigger a volcano that is nearly ready to erupt anyway. Similarly, Mulargia et al. (1992) used a statistical nonparametric pattern-recognition algorithm to examine possible precursors of eruptions at Mount Etna 1974–1990, including the number of earthquakes in five surrounding tectonic regions and their maximum magnitude. An 80-day interval straddling the eruption onset was considered, and the pattern of 12 factors in each interval noted. The only statistically significant result was that flank eruptions were linked to seismicity with more than six earthquakes in the surrounding regions during the 80-day period. All 11 of the flank eruptions considered satisfied this condition, but only 11 out of 24 (in a total of 62) windows with this level of seismicity corresponded to earthquakes. Summit eruptions possessed no discernable pattern.

Mulargia (1992) used the intensity cross product for two series of events at times $0 < s_1 < \dots < s_{M(T)} < T$ and $0 < t_1 < \dots < t_{N(T)} < T$. For a lag u , this is estimated by

$$\hat{p}_{MN}(u) = \frac{n(u, h)}{2hT}, \quad (56)$$

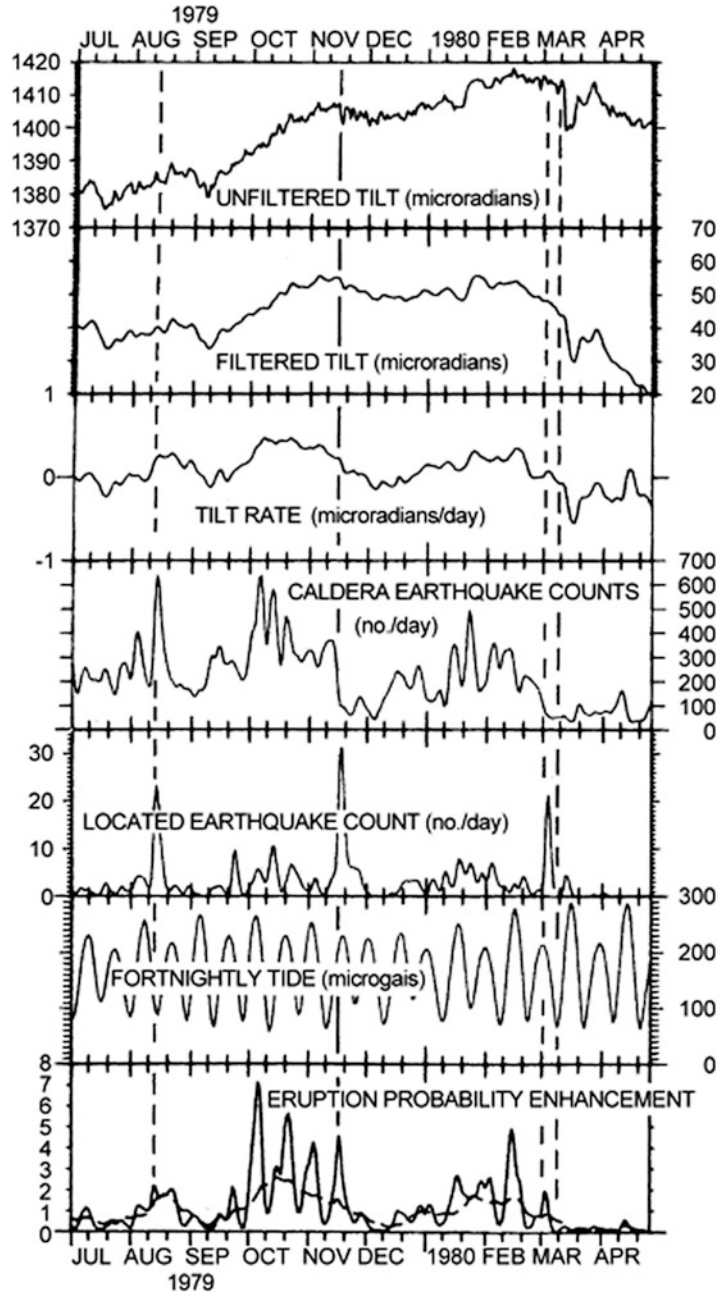
where $n(u, h) = \# \{i : t_j + u - h \leq s_i \leq t_j + u + h, \text{ for some } j\}$ is the number of events from the process M that occur within a window of length $2h$ of the lag of a point in the process N . Under a null hypothesis of two independent Poisson processes, the significance level can be calculated as

$$P = \sum_{k=n(u, h)}^{N(T)} \binom{k}{N(T)} \left(\frac{2hM(T)}{T} \right)^k. \quad (57)$$

This was applied to the flank eruptive activity and local seismicity at Mt Etna 1974–1990. The data

Volcanic Eruptions: Stochastic Models of Occurrence Patterns,

Fig. 24 Kilauea volcano: several parameters and eruption probability enhancement at an expanded timescale. The figure spans 10 months during 1979–1980 and includes the November 16, 1979, eruption (*solid vertical line*) and intrusions on August 12, 1979, and March 2 and 10, 1980 (*dashed vertical lines*). Intrusions, like eruptions, are generally accompanied by summit deflation, an earthquake swarm and harmonic tremor, but magma fails to reach the surface. The eruption probability enhancement is plotted as a 1-day outlook (*solid line*) and 30-day outlook (*dashed line*). The 1-day probability depends on the fortnightly tide and oscillates with it. The 30-day probability is a function of smoothed tilt and earthquake parameters and does not depend on the tide. Note that eruption probability enhancement was higher than average (greater than 1) for about 2 months prior to the November 1979 eruption and March 1980 intrusions (From Klein FW (1984) Eruption forecasting at Kilauea Volcano, Hawaii. J Geophys Res 89:3064 Fig. 4 © 1984 by the American Geophysical Union)



consisted of 11 eruptions and 12 seismic sequences (Gasperini et al. 1990). Using a window of length $2h = 10$ days, Mulargia (1992) found a highly significant association peak at a lag u between -7 and 3 days. This indicates that local earthquakes and flank eruptions are largely concomitant phenomena.

Marzocchi et al. (1993) examined the correlation between eruptions of Vesuvius and earthquakes of $M > 5.4$ in the southern Apennines, Calabrian Arc, and Sicily, 1631–1944. The eruptions and earthquakes were binned into intervals of 1, 2, 3, and 4 years, and the Spearman rank correlation

$$\rho_{XY}(k) = \frac{\sum_{i=1}^m R(X_{i+k})R(Y_i) - m(m+1)^2/4}{\left(\sum_{i=1}^m R(X_{i+k})^2 - m(m+1)^2/4\right)^2 \cdot \left(\sum_{i=1}^m R(Y_i)^2 - m(m+1)^2/4\right)^2} \quad (58)$$

calculated, where X_i, Y_i are the total seismic moment and number of eruptions in the i th time bin; $R(\cdot)$ is the rank, from smallest to largest, of the quantities; and k is the lag. Varying this lag produces the Spearman correlogram. Marzocchi et al. (1993) observed changes in eruptive activity following increases in seismicity in the southern Apennines after 6–13 years and with delays of 27–30 years and 36–39 years for earthquakes in Sicily and the Calabrian Arc, respectively. Nostro et al. (1998) investigated this coupling (cf. Fig. 25) in terms of the Coulomb stress change on a magma body beneath Vesuvius caused by a slip on Apennine normal faults (promoting eruptions by earthquakes) and on fault planes caused by voiding a buried magma body (promoting earthquakes by eruptions), as shown in Fig. 26. Using these results to select earthquakes and eruptions improved the significance of the Marzocchi et al. (1993) analysis.

Cardaci et al. (1993) used a cross correlation technique to investigate the relation between seismic and volcanic data at Mount Etna 1975–1987. The seismic and eruptive data were averaged over seven-day intervals, with the cross correlation being

$$\rho_{XY}(x) = \frac{\sigma_{XY}(k)}{\sqrt{\sigma_X^2 \sigma_Y^2}} \quad (59)$$

for $k = 0, \pm 1, \dots, \pm K$, where

$$\sigma_{XY}(k) = \begin{cases} (1/n) \sum_{i=1}^{n-k} (X_i - \mu_X)(Y_{i+k} - \mu_Y), & k = 0, 1, \dots, K \\ (1/n) \sum_{i=1-k}^n (X_i - \mu_X)(Y_{i+k} - \mu_Y), & k = -1, \dots, -K \end{cases} \quad (60)$$

is the cross covariance function and μ_X, μ_Y and σ_X^2, σ_Y^2 are the means and variances, respectively, of the two time series $X_1, \dots, X_n$ and $Y_1, \dots, Y_n$. The significance of the cross correlation can be assessed from the fact that $\rho_{XY}(k) \times$

$\sqrt{(m-2)/(1-\rho_{XY}(k))}$ is asymptotically distributed as a t -distribution with $m-2$ degrees of freedom, where m is the number of overlapped positions between the two series. Cardaci et al. (1993) arrived at a similar conclusion to Mulargia et al. (1992), identifying low-frequency events, whose rate of occurrence increases from 17 to 108 days prior to flank eruptions, as the best precursor. Summit events had a less clear relation with volcanic tremor rather than seismic activity.

Linde and Sacks (1998) examined the historical record of eruptions ($\text{VEI} \geq 2$) and large earthquakes, to see whether there are more of the former within 2 days of the latter. For earthquakes with ($M \geq 8$), there are a significant excess of eruptions within 750 km of the epicenter, while for earthquakes with $7.0 \leq M \leq 7.9$, the effect is only observed within a radius of 200 km. In all, 20 such earthquake-volcano pairs are observed 1587–1974. In addition, Linde and Sacks (1998) hypothesized that eruptions in unison of volcanoes separated by hundreds of kilometers could be due to the second eruption being triggered by earthquakes associated with the first. Marzocchi (2002) elaborated on this analysis by defining a perturbation function

$$\phi_i^{(k)}(\Delta) = \sum_j M_j w(d_{jk}) \mathbf{I}_{((i-1)\Delta, i\Delta)}(t_k - s_j), \quad (61)$$

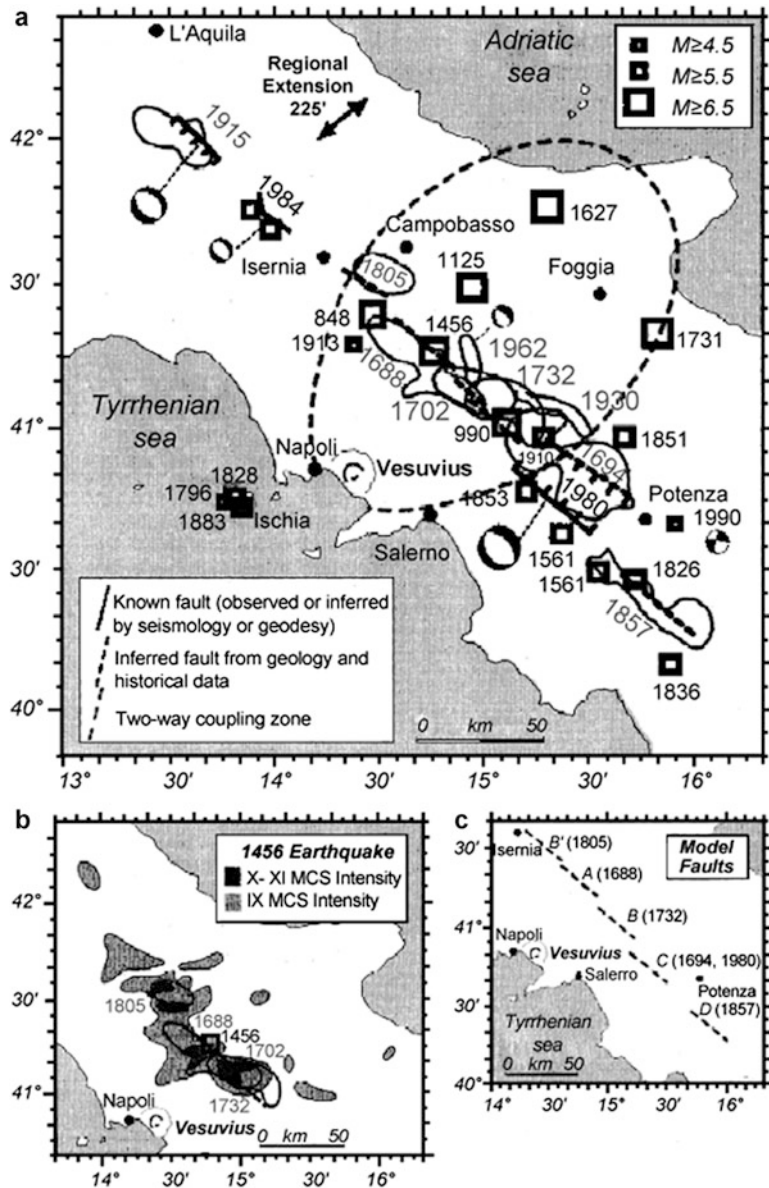
where

$$\mathbf{I}_A(x) = \begin{cases} 1 & x \in A \\ 0 & x \notin A \end{cases} \quad (62)$$

is the indicator function, s_j and M_j are the time and seismic moment of the j th earthquake, t_k is the onset time of the k th eruption, and i is the

**Volcanic Eruptions:
Stochastic Models of
Occurrence Patterns,**

Fig. 25 (a) Sites of historical earthquakes in south central Italy, associated faults and Vesuvius volcano. Epicenters or, if unknown, isoseismals of modified Mercalli intensity are shown. The coupling zone is the region in which normal-faulting earthquakes may promote Vesuvius eruptions and vice versa. (b) Isoseismals for the 1456 earthquake with 1688–1805 isoseismal zones. (c) Simplified model faults A–D used for calculations, with the dates of earthquakes associated with each fault segment (From Nostro C, Stein RS, Cocco M, Belardinelli ME, Marzocchi W (1998) Two-way coupling between Vesuvius eruptions and southern Apennine earthquakes, Italy, by elastic stress transfer. *J Geophys Res* 103:24488, Fig. 1 © 1998 by the American Geophysical Union)

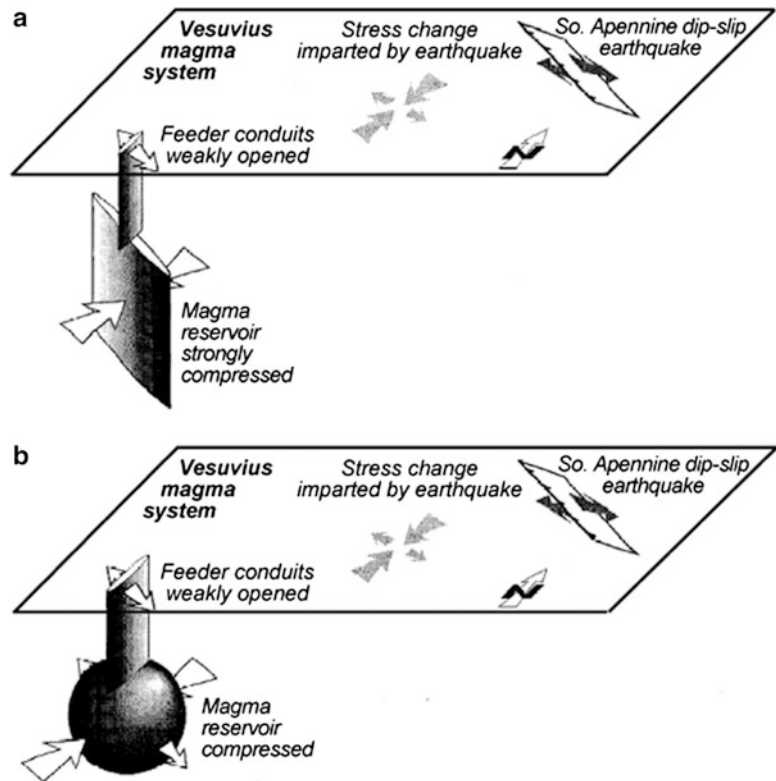


lag. The distance between the epicenter of the j th earthquake and the k th eruption is d_{jk} , which is weighted by a (sharply) decreasing function $w(\cdot)$. The statistical significance of the calculated values is assessed by a randomization test. The idea is that the perturbation on the volcano by earthquakes is proportional to the seismic moment and decays with distance. Of the eight eruptions with $VEI \geq 5$ examined, using $\Delta =$

3 and 5 years, five have a significant peak at 0–5 years in the perturbation function, calculated from earthquakes with $M \geq 7$ (792 events, 1900–1999). Using an earthquake catalog with $M \geq 6$ (3,186 events, 1950–1997), two of four events have a significant peak at 30–35 years before the eruption. Marzocchi et al. (2004a) expanded this analysis to the 62 twentieth-century eruptions with $VEI \geq 4$, identifying

Volcanic Eruptions: Stochastic Models of Occurrence Patterns,

Fig. 26 Schematic illustration of the response of a hypothetical Vesuvius magmatic system to a southern Apennine normal-faulting earthquake for (a) a buried dike in which at least one fissure or feeder conduit strikes NE and (b) a buried spherical magma chamber with a NE striking fissure (From Nostro C, Stein RS, Cocco M, Belardinelli ME, Marzocchi W (1998) Two-way coupling between Vesuvius eruptions and southern Apennine earthquakes, Italy, by elastic stress transfer. *J Geophys Res* 103:24497, Fig. 3 © 1998 by the American Geophysical Union)



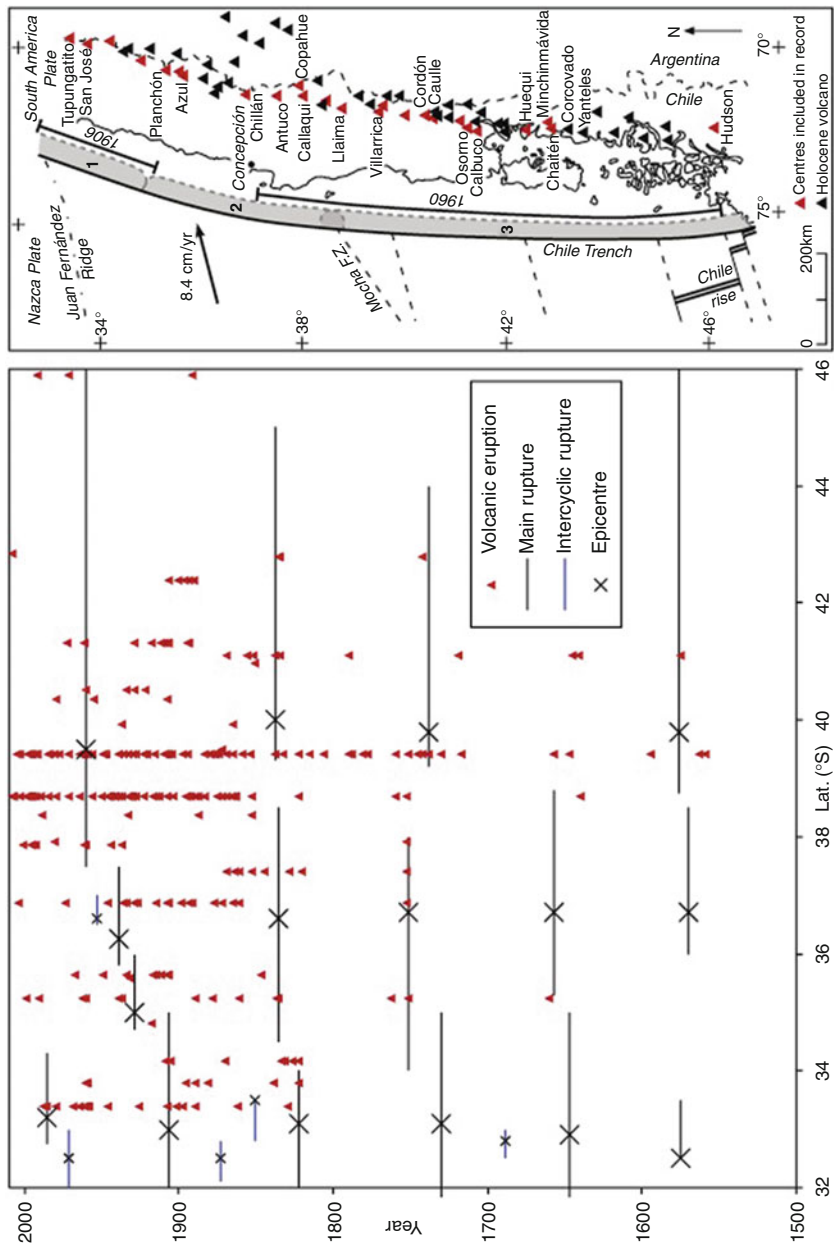
peaks at 0–5 and 20–35 years. The perturbations for $\text{VEI} \geq 4$ eruptions were shown to be significantly larger than those for $\text{VEI} = 1$ and $\text{VEI} = 2$ eruptions. This suggests that the likelihood of large explosive eruptions is affected by post-seismic stress variations induced by large earthquakes up to decades earlier and hundreds of kilometers distant. No spatiotemporal clustering of the earthquakes and eruptions was detected.

Watt et al. (2009) showed that the time-smoothed eruption rate along the Andean southern volcanic zone 1850-present (Fig. 27) possessed two significant (against a Poisson process) peaks in 1906 and 1960, immediately following the two largest ($M_W 9.5$ and $M_W 8.3$, respectively) earthquakes in the Chilean subduction zone. However, there was no significant effect from the 1939 and 1985 $M_W 8.0$ earthquakes, and only slight evidence of triggering following the 1928 $M_W 8.2$ earthquake. The

possible explanations include the spatial extent of the events and the time intervals between them. These factors were examined on Indonesian data (Fig. 28) by Bebbington and Marzocchi (2011) using earthquakes as a “triggering term” (a time-decreasing multiplicative factor on the hazard) in a Poisson process, Weibull renewal process (Eq. 16) or a volume-history model (Eq. 29). Evaluating the significance of additional parameters in a more complicated model using AIC (Akaike 1977), significant triggering was observed for some volcanoes, but only under the volume-history model (Fig. 29), implying that the effect is one of “clock advance” – bringing forward the time of the next eruption.

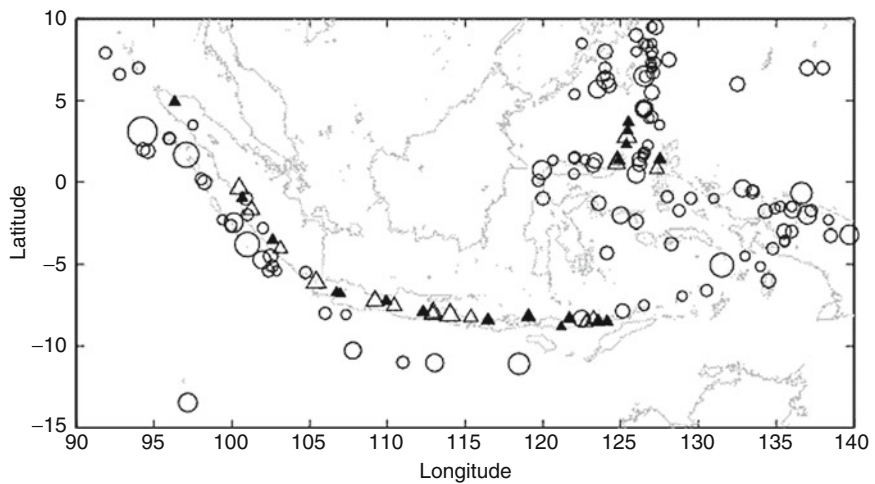
Model Assessment

A direction only now receiving attention is the validation of forecasting models, against data not



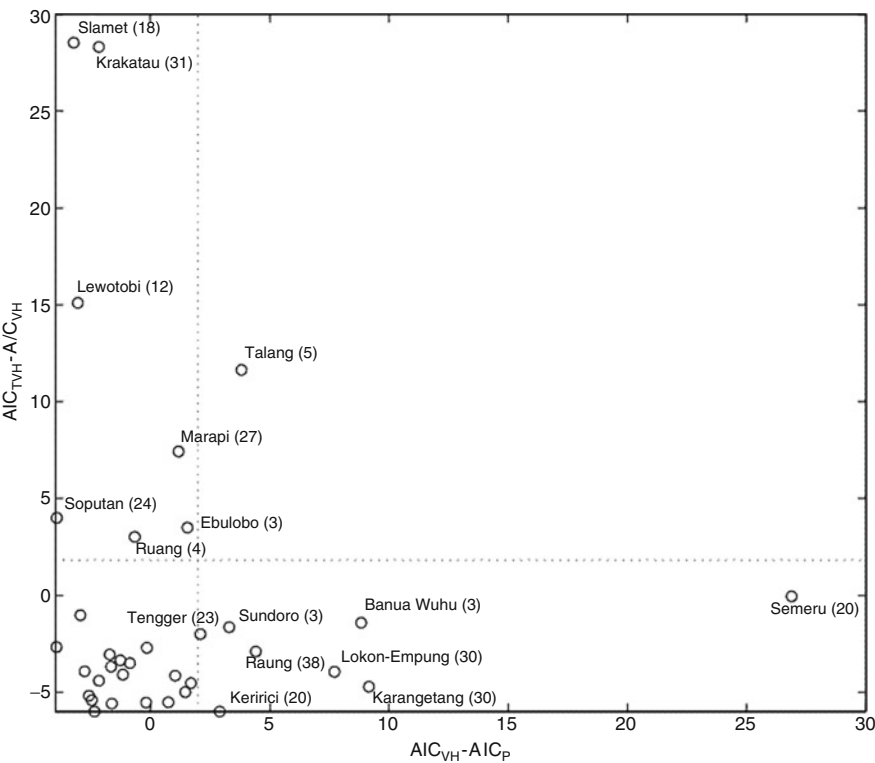
rupture zone segments are shown with approximate limits, corresponding to Central Chile (1), Central Valley (2), and Southern Nazca (3) rupture zones. The cause of the boundary between (1) and (2) is unclear, while the (2) and (3) boundary may be related to subduction of the Mocha fracture zone. Subduction of the Juan Fernandez seamounts may be related to the onset of the volcanic gap north of the SVZ (Reprinted from Watt SFL, Pyle DM, Mather TA (2009) The influence of great earthquakes on volcanic eruption rate along the Chilean subduction zone. Earth Planet Sci Lett 277:401, © 2008 Elsevier B.V., with permission from Elsevier)

Volcanic Eruptions: Stochastic Models of Occurrence Patterns,
Fig. 27 Historically recorded large earthquakes ($M_w > 7.5$) in central and southern Chile, showing main ruptures ($M_w > 8$) and inter-cyclic events, with epicenters and approximate rupture lengths. Earthquake occurrence shows a cyclic temporal clustering and broad southward stepping pattern. Reliably recorded historic volcanic eruptions are depicted, showing that a few centers dominate the record and that eruptions in the far south of the region appear to be underrepresented. The map depicts the regional tectonics, with volcano locations. Named volcanoes are discussed in the text. Schematic



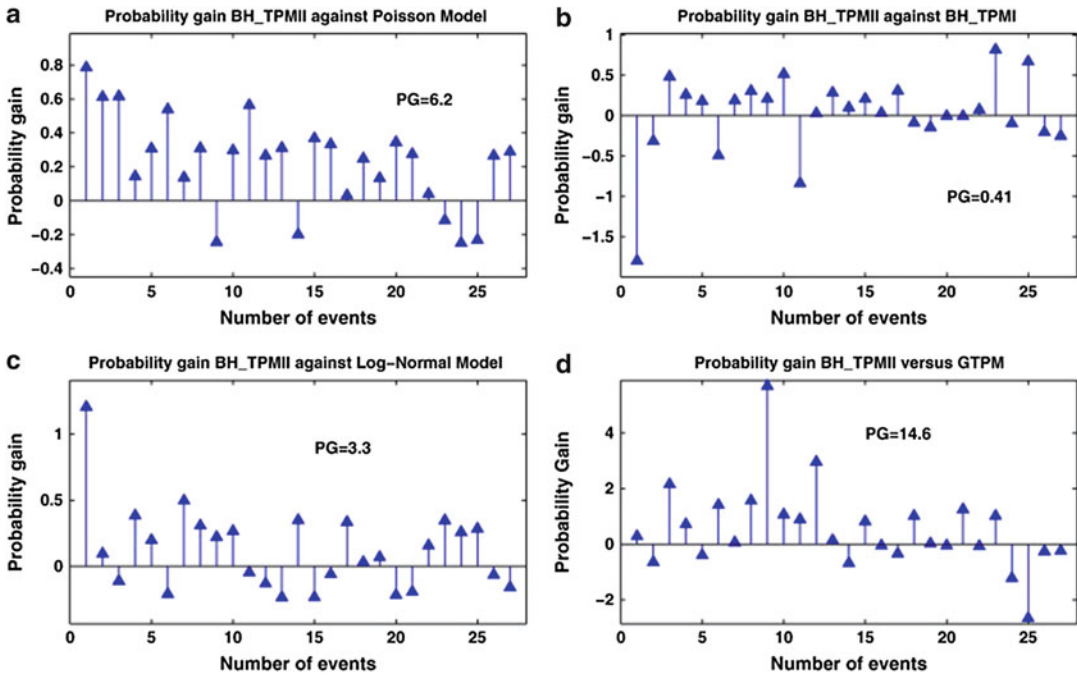
Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 28 Earthquakes (circles) and volcanoes (open triangles for at least 12 eruptions, solid triangles for between 3 and 9 eruptions, all VEI ≥ 2). The symbol size scales as the magnitude of the earthquake (minimum 7.0, maximum 9.0) and the log of the number of eruptions

(minimum 3, maximum 38), respectively (From Bebbington MS, Marzocchi W (2011) Stochastic models for earthquake triggering of volcanic eruptions. J Geophys Res 116:B05204, Fig. 1 © 2011 by the American Geophysical Union)



Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 29 AIC improvement from incorporating volume history (horizontal axis) and then earthquake triggering (vertical axis). The dotted lines indicate the 5 % level of significance. The number of eruptions follows the

volcano name (From Bebbington MS, Marzocchi W (2011) Stochastic models for earthquake triggering of volcanic eruptions. J Geophys Res 116:B05204, Fig. 6 © 2011 by the American Geophysical Union)

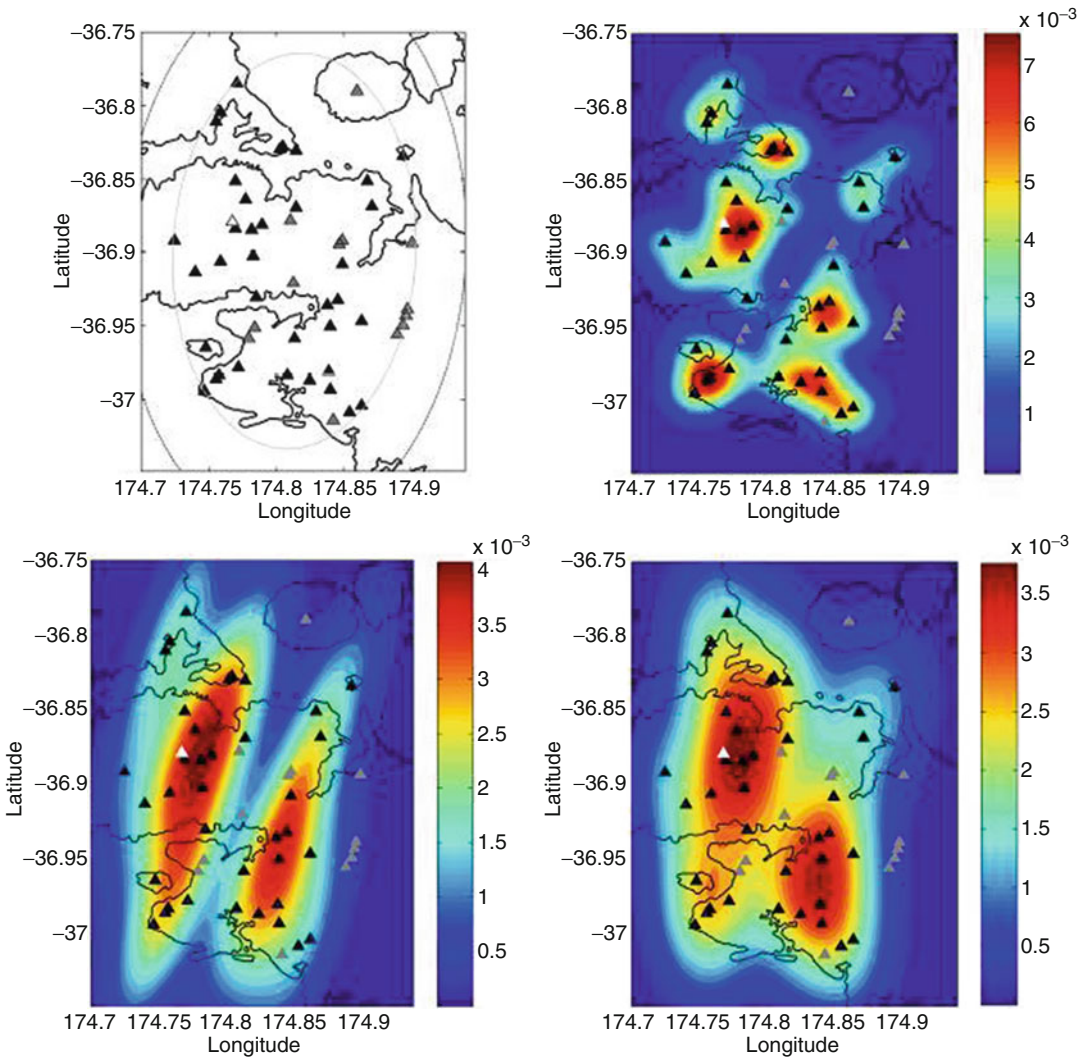


Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 30 “Punctual probability gain” of the BH TPMII for each event after the learning phase against: in panel (a) Poisson model (Klein 1982), in panel (b) BH TPM (Passarelli et al. 2010a), in panel (c) lognormal model (Bebington and Lai 1996b), and in panel (d) generalized time-predictable model (Sandri et al. 2005). Values greater than zero indicate when BH TPM model performs better forecast than the reference models. Positive

values indicate that BH TPM has better forecasting ability than the alternative model. Global probability gains are reported as “PG” in each of the four cases (Reprinted from Passarelli L, Sanso B, Sandri L, Marzocchi W (2010) Testing forecasts of a new Bayesian time-predictable model of eruption occurrence. *J Volcanol Geotherm Res* 198:66, © 2010 Elsevier B.V., with permission from Elsevier)

used to fit them. This is a different question from goodness of fit, which is assessed against the data used to fit the model. The difference lies in how much of the model fit is “signal” versus the amount of “noise.” In general, more complex models (those with a greater number of parameters) tend to improve their goodness of fit scores by “overfitting” this random noise. When used prospectively, the resulting forecast maps this noise onto the forecast, which can have a very different random element. Ideally this is dealt with by dividing the data into two or three parts. Candidate models are *fitted* on the earliest part of the data, the best model is selected according to their performance on the second part of the data, and the chosen model accuracy is assessed (*validation*) on the final part. If only considering one model, then the second step is skipped.

Unfortunately, typical volcanic records are too short to permit of this procedure. By considering the probability gain, or improvement in the log likelihood of forecast events, Passarelli et al. (2010b) successfully validated models for Kilauea in this manner (Fig. 30), aided by the fact that the volcano has been (at least approximately) in a steady state for most of the twentieth century. However, many perhaps even a majority (Bebington 2010) of volcanoes have activity that waxes and wanes over timescales of a century or more. Hence, the fitting and validation data will be qualitatively different. A fundamental limitation is that a statistical model should not be able to forecast behavior that is not in the fitting data, although a number of models with constantly increasing rate have been proposed (Ho 1992a; Salvi et al. 2006; Smethurst et al. 2009). The



Volcanic Eruptions: Stochastic Models of Occurrence Patterns, Fig. 31 Spatial hazard estimates. The white triangle (Mt Eden) is being forecast using the age and location of the previous centers (*black triangles*). The *gray triangles* indicate eruptions that have not yet occurred. *Top left*: Inner and outer ellipses, *Top right*: Gaussian isotropic kernel (LSCV), *Bottom left*: Gaussian

anisotropic kernel (LSCV), *Bottom right*: Gaussian anisotropic kernel (SAMSE) (Reprinted (and colored) from Bebbington MS (2013) Assessing spatio-temporal eruption forecasts in a monogenetic volcanic field. *J Volcanol Geotherm Res* 252:20, © 2013 Elsevier B.V., with permission from Elsevier)

recommended approach (Bebbington 2013) to dealing with this problem is to approximate the validation step using a criterion such as the Akaike Information Criterion (AIC) (Akaike 1977) which penalizes additional parameters. AIC appears to be more suitable than BIC for typical volcanic records (Bebbington 2007). Note that the corrected AIC (Hurvich and Tsai

1989) should not be used for the class of point process models, as it has only been proven to be valid for linear regression and autoregressive models (Claeskens and Hjort 2008).

There are a few exceptions to the problems with short record lengths. In the case of geological data, event times are known only to a probability distribution. Hence, it is possible to repeatedly

simulate independent realizations of the event time order (Condit and Connor 1996). Hence, even short records can be divided into fitting and validation data, the brevity of the data being partially compensated for by repetition. This can be coupled with sequential forecasting through each realization, as used, for example, in (Bebbington 2013) to distinguish between possible spatiotemporal models for the Auckland Volcanic Field (Fig. 31). The probability gains (Bebbington 2013; Passarelli et al. 2010b) are an empirical approach to assessing the effectiveness of a given forecasting model. The performance can be bounded above by the information gain (Vere-Jones 1998), which can be calculated for certain renewal and volume-dependent hazard models (Bebbington 2005; Imoto 2004).

Future Directions

One exciting area is the incorporation in models of ancillary data, i.e., data beyond the time and/or location of the event. Beyond the incorporation of eruptive volume in the family of time-predictable models, this has recently been accomplished for earthquake triggering (Bebbington and Marzocchi 2011), dike emplacement (Kiyosugi et al. 2012), and geochemical forcing in both time (Green et al. 2013) and space (Le Corvec et al. 2013); the possibilities are limited only by imagination and the availability of data.

Another area of development is that of accounting for measurement error and missing observations. Geological dating provides age estimates as means and standard deviations, and standard practice is to use Monte Carlo simulation, repeatedly drawing samples from the appropriate distributions. This is complicated when stratigraphic constraints are added. A similar approach can be taken when onset dates are incompletely specified, although a more common alternative is to use the center of the given time window. Missing events can be implicitly incorporated in a model by using robust estimation techniques (Wang and Bebbington 2013) or explicitly by accounting for them in a hidden Markov model framework (Wang and Bebbington 2012), where

the hidden state represents the number of missing onsets between observed onsets. These methods have so far only been developed for the Weibull process and the renewal process, respectively.

More generally, much of the relevant literature on stochastic processes has not been applied to problems in volcanology. In particular, (marked) point processes, alternating renewal processes, and semi-Markov processes would seem to be natural candidates for modeling volcanic activity. A number of techniques from time-series analysis (see Young (2006) for a brief review) may prove useful for the investigation of volcanic data at various timescales.

In the application of stochastic models, and the understanding to be gleaned from them, volcanology lags well behind seismology. This is despite the inherent advantages of a phenomenon in which spatial degrees of freedom can often be ignored. An obvious future direction is the borrowing of more techniques from seismology. Unlike seismology, a certain amount of effort (and money) can often produce additional data through coring and dating. This is a great advantage in statistical terms, especially as the data can be “targeted” to some degree as required by the model. This additional data makes it possible in volcanology to detect, and forecast, nonstationary behavior and to relate this to physical changes in the volcano (Turner et al. 2008b).

Although this entry has concentrated on models for occurrence patterns, actual volcanic hazard is dependent on the nature and behavior of volcanic products such as lava, ash fall, lahars, and pyroclastic flows. Incorporating existing sophisticated numerical models of such behavior with stochastic models of occurrence, location/direction, and magnitude will allow more accurate estimation of the hazard to life and property.

Bibliography

Primary Literature

- Alvarez-Ramirez J, Sosa E, Hernandez-Martinez E (2011) In-phase dynamics of the exhalation sequence in Popocatepetl volcano and slow-slip events in Cocos-North American plate boundary. *J Volcanol Geotherm Res* 200:83–90

- Akaike H (1977) On entropy maximization principle. In: Krishnaiah PR (ed) *Applications of statistics*. North-Holland, Amsterdam, pp 27–41
- Bacon CR (1982) Time-predictable bimodal volcanism in the Coso range, California. *Geology* 10:65–69
- Bain LJ (1978) *Statistical analysis of reliability and life testing models*. Marcel Dekker, New York
- Bebbington MS (2005) Information gains for stress release models. *Pure Appl Geophys* 162:2299–2319
- Bebbington MS (2007) Identifying volcanic regimes using hidden Markov models. *Geophys J Int* 171:921–942
- Bebbington M (2008) Incorporating the eruptive history in a stochastic model for volcanic eruptions. *J Volcanol Geotherm Res* 175:325–333
- Bebbington M (2010) Trends and clustering in the onsets of volcanic eruptions. *J Geophys Res* 115, B01203
- Bebbington MS (2013) Assessing spatio-temporal eruption forecasts in a monogenetic volcanic field. *J Volcanol Geotherm Res* 252:14–28
- Bebbington M, Cronin SJ (2011) Spatio-temporal hazard estimation in the Auckland Volcanic Field, New Zealand, with a new event-order model. *Bull Volcanol* 73:55–72
- Bebbington MS, Lai CD (1996a) On nonhomogeneous models for volcanic eruptions. *Math Geol* 28:585–600
- Bebbington MS, Lai CD (1996b) Statistical analysis of New Zealand volcanic occurrence data. *J Volcanol Geotherm Res* 74:101–110
- Bebbington MS, Lai CD (1998) A generalised negative binomial and applications. *Commun Stat Theor Methods* 27:2515–2533
- Bebbington MS, Marzocchi W (2011) Stochastic models for earthquake triggering of volcanic eruptions. *J Geophys Res* 116, B05204
- Behncke B, Neri M (2003) Cycles and trends in the recent eruptive behaviour of Mount Etna (Italy). *Can J Earth Sci* 40:1405–1411
- Behncke B, Neri M, Nagay A (2005) Lava flow hazard at Mount Etna (Italy): new data from a GIS-based study. *Geol Soc Am, Special Paper* 396:189–208
- Box GEP, Draper NR (1987) *Empirical model-building and response surfaces*. Wiley, New York.
- Branca S, Del Carlo P (2004) Eruptions of Mt. Etna during the past 3,200 years: a revised compilation integrating the historical and stratigraphic records. In: Bonaccorso A, Calvari S, Coltelli M, Del Negro C, Falsaperla S (eds) *Mt. Etna: volcano laboratory*, vol 143, *Geophysical monograph series*. American Geophysical Union, Washington, DC, pp 1–27
- Bronk Ramsey C (1995) Radiocarbon calibration and analysis of stratigraphy: the OxCal program. *Radiocarbon* 37:425–430
- Burt ML, Wadge G, Curnow RN (2001) An objective method for mapping hazardous flow deposits from the stratigraphic record of stratovolcanoes: a case example from Montagne Pelee. *Bull Volcanol* 63:98–111
- Burt ML, Wadge G, Scott WA (1994) Simple stochastic modelling of the eruption history of a basaltic volcano: Nyamuragira, Zaire. *Bull Volcanol* 56:87–97
- Cappello A, Neri M, Acocella V, Gallo G, Vicari A, Del Negro C (2012) Spatial vent opening probability map of Etna volcano (Sicily, Italy). *Bull Volcanol* 74: 2083–2094
- Cardaci C, Falsaperla S, Gasperini P, Lombardo G, Marzocchi W, Mulargia F (1993) Cross-correlation analysis of seismic and volcanic data at Mt Etna volcano, Italy. *Bull Volcanol* 55:596–603
- Carta S, Figari R, Sartoris G, Sassi R, Scandone R (1981) A statistical model for Vesuvius and its volcanological implications. *Bull Volcanol* 44:129–151
- Casetti G, Frazzetta G, Romano R (1981) A statistical analysis in time of the eruptive events on Mount Etna (Italy) from 1323 to 1980. *J Volcanol Geotherm Res* 44: 283–294
- Claeskens G, Hjort NL (2008) *Model selection and model averaging*. Cambridge University Press, Cambridge
- Coleman NM, Abramson LR, Marsh BD (2004) Testing claims about volcanic disruption of a potential geologic repository at Yucca Mountain, Nevada. *Geophys Res Lett* 31, L24601
- Condit CD, Connor CB (1996) Recurrence rates of volcanism in basaltic volcanic fields: an example from the Springerville volcanic field, Arizona. *Geol Soc Am Bull* 108:1225–1241
- Connor CB (1990) Cinder cone clustering in the Trans-Mexican volcanic belt: implications for structural and petrologic models. *J Geophys Res* 95:19395–19405
- Connor CB, Connor LJ (2009) Estimating spatial density with kernel methods. In: Connor CB, Chapman NA, Connor LJ (eds) *Volcanic and tectonic hazard assessment for nuclear facilities*. Cambridge University Press, Cambridge, pp 346–368
- Connor CB, Hill BE (1995) Three nonhomogeneous Poisson models for the probability of basaltic volcanism: application to the Yucca Mountain region, Nevada. *J Geophys Res* 100:10107–10125
- Connor CB, Stamatakis JA, Ferrill DA, Hill BE, Ofoegbu GI, Conway FM, Sagar B, Trapp J (2000) Geologic factors controlling patterns of small-volume basaltic volcanism: application to a volcanic hazards assessment at Yucca Mountain, Nevada. *J Geophys Res* 105:417–432
- Connor CB, Sparks RSJ, Mason RM, Bonadonna C, Young SR (2003) Exploring links between physical and probabilistic models of volcanic eruptions: the Soufriere Hills volcano, Montserrat. *Geophys Res Lett* 30:1701
- Conway FM, Connor CB, Hill BE, Condit CD, Mullaney K, Hall CM (1998) Recurrence rates of basaltic volcanism in SP cluster, San Francisco volcanic field, Arizona. *Geology* 26:655–658
- Crandell DR, Mullineaux DR, Rubin M (1975) Mount St. Helens volcano: recent and future behaviour. *Science* 187:438–441
- Cronin S, Bebbington M, Lai CD (2001) A probabilistic assessment of eruption recurrence on Taveuni volcano, Fiji. *Bull Volcanol* 63:274–288

- Croweller HS, Arora A, Brown SK, Cottrell E, Deligne NI, Guerrero NO, Hobbs L, Kiyosugi K, Loughlin SC, Lowndes J, Nayembil M, Siebert L, Sparks RSJ, Takarada S, Venzke E (2012) Global database on large magnitude explosive volcanic eruptions (LaMEVE). *J Appl Volcanol* 1:4
- Crowe BM, Johnson ME, Beckman RJ (1982) Calculation of the probability of volcanic disruption of a high-level radioactive waste repository within southern Nevada, USA. *Radioact Waste Manag* 3:167–190
- Crowe BM, Wallmann P, Bowker LM (1998) Probabilistic modeling of volcanism data: final volcanic hazard studies for the Yucca Mountain site. In: Perry FV et al (eds) *Volcanism studies: final report for the Yucca Mountain project*. Los Alamos National Laboratory Report LA-13478. Los Alamos National Laboratory, Los Alamos, 415 pp
- Decker RW (1986) Forecasting volcanic eruptions. *Ann Rev Earth Planet Sci* 14:267–291
- De la Cruz-Reyna S (1991) Poisson-distributed patterns of explosive eruptive activity. *Bull Volcanol* 54:57–67
- De la Cruz-Reyna S (1993) Random patterns of occurrence of explosive eruptions at Colima volcano, Mexico. *J Volcanol Geotherm Res* 55:51–68
- De la Cruz-Reyna S, Carrasco-Nunez G (2002) Probabilistic hazard analysis of Citlaltepētēl (Pico de Orizaba) volcano, eastern Mexican volcanic belt. *J Volcanol Geotherm Res* 113:307–318
- Deligne NI, Coles SG, Sparks RSJ (2010) Recurrence rates of large explosive volcanic eruptions. *J Geophys Res* 115, B06203
- Duong T (2007) ks: kernel density estimations and kernel discriminant analysis for multivariate data in R. *J Stat Softw* 21:1–16
- Duong T, Hazelton ML (2003) Plug-in bandwidth selectors for bivariate kernel density estimation. *J Nonparametr Stat* 15:17–30
- Dubois J, Cheminee JL (1991) Fractal analysis of eruptive activity of some basaltic volcanoes. *J Volcanol Geotherm Res* 45:197–208
- Eliasson J, Larsen G, Gudmundsson MT, Sigmundsson F (2006) Probabilistic model for eruptions and associated flood events in the Katla caldera, Iceland. *Comput Geosci* 10:179–200
- Felpeo A, Marti J, Ortiz R (2007) A GIS-based automatic system for volcanic hazard assessment and risk management. *J Volcanol Geotherm Res* 166:106–116
- Feuillet N, Cocco M, Musumeci C, Nostro C (2006) Stress interaction between seismic and volcanic activity at Mt Etna. *Geophys J Int* 164:697–718
- Furlan C (2010) Extreme value methods for modelling historical series of large volcanic magnitudes. *Stat Model* 10:113–132
- Garcia-Aristizabal A, Marzocchi W, Fujita E (2012) A Brownian model for recurrent volcanic eruptions: an application to Miyakejima volcano (Japan). *Bull Volcanol* 74:545–558
- Garcia-Aristizabal A, Selva J, Fujita E (2013) Integration of stochastic models for long-term eruption forecasting into a Bayesian event tree scheme: a basis method to estimate the probability of volcanic unrest. *Bull Volcanol* 75:689
- Gasperini P, Gresta S, Mulargia F (1990) Statistical analysis of seismic and eruptive activities at Mt. Etna during 1978–1987. *J Volcanol Geotherm Res* 40:317–325
- Geomatrix Consultants (1996) Probabilistic volcanic hazard analysis for Yucca Mountain, Nevada. Report BA0000000-1717-220-00082, Geomatrix Consultants, San Francisco
- Geyer A, Marti J (2008) The new worldwide Collapse Caldera Database (CCDB): a tool for studying and understanding caldera processes. *J Volcanol Geotherm Res* 175:334–354
- Green R, Bebbington MS, Cronin SJ, Jones G (2013) Geochemical precursors for eruption repose length. *Geophys J Int* 193:855–873
- Godano C, Civetta L (1996) Multifractal analysis of Vesuvius volcano eruptions. *Geophys Res Lett* 23:1167–1170
- Gusev AA (2008) Temporal structure of the global sequence of volcanic eruptions: order clustering and intermittent discharge rate. *Phys Earth Planet Int* 166:203–218
- Gusev AA, Ponomareva VV, Braitseva OA, Melekestsev IV, Sulerzhitsky LD (2003) Great explosive eruptions on Kamchatka during the last 10,000 years: self-similar irregularity of the output of volcanic products. *J Geophys Res* 108(B2):2126
- Guttorp P, Thompson ML (1991) Estimating second-order parameters of volcanicity from historical data. *J Am Stat Assoc* 86:578–583
- Hill BE, Connor CB, Jarzempa MS, La Femina PC, Navarro M, Strauch W (1998) 1995 eruptions of Cerro Negro volcano, Nicaragua, and risk assessment for future eruptions. *Geol Soc Am Bull* 110:1231–1241
- Ho CH (1990) Bayesian analysis of volcanic eruptions. *J Volcanol Geotherm Res* 43:91–98
- Ho CH (1991) Nonhomogeneous Poisson model for volcanic eruptions. *Math Geol* 23:167–173
- Ho CH (1992a) Risk assessment for the Yucca Mountain high-level nuclear waste repository site: estimation of volcanic disruption. *Math Geol* 24:347–364
- Ho CH (1992b) Statistical control chart for regime identification in volcanic time-series. *Math Geol* 24:775–787
- Ho CH (1995) Sensitivity in volcanic hazard assessment for the Yucca Mountain high-level nuclear waste repository site: the model and the data. *Math Geol* 27:239–258
- Ho CH, Smith EI (1997) Volcanic hazard assessment incorporating expert knowledge: application to the Yucca Mountain region, Nevada, USA. *Math Geol* 29:615–627
- Ho CH, Smith EI (1998) A spatial-temporal/3-D model for volcanic hazard assessment: application to the Yucca Mountain region, Nevada. *Math Geol* 30:497–510
- Ho CH, Smith EI, Feuerbach DL, Naumann TR (1991) Eruptive probability calculation for the Yucca Mountain site, USA: statistical estimation of recurrence rates. *Bull Volcanol* 54:50–56

- Hurvich CM, Tsai CL (1989) Regression and time series model selection in small samples. *Biometrika* 76: 297–307
- Imoto M (2004) Probability gains expected for renewal process models. *Earth Planets Space* 56:563–571
- Jaquet O, Lantuejoul C, Goto J (2012) Probabilistic estimation of long-term volcanic hazard with assimilation of geophysics and tectonic data. *J Volcanol Geotherm Res* 235–236:29–36
- Jaquet O, Low S, Martinelli B, Dietrich V, Gilby D (2000) Estimation of volcanic hazards based on Cox stochastic processes. *Phys Chem Earth (A)* 25:571–579
- Jupp TE, Pyle DM, Mason BG, Dade WB (2004) A statistical model for the timing of earthquakes and volcanic eruptions influenced by periodic processes. *J Geophys Res* 109, B02206
- Kiyosugi K, Connor CB, Zhao D, Connor LJ, Tanaka K (2010) Relationships between volcano distribution, crustal structure, and P-wave tomography: an example from the Abu Monogenetic Volcano Group, SW Japan. *Bull Volcanol* 72:331–340
- Kiyosugi K, Connor CB, Wetmore PH, Ferwerda BP, Germa AM, Connor LJ, Hintz AR (2012) Relationship between dike and volcanic conduit distribution in a highly eroded monogenetic volcanic field: San Rafael, Utah, USA. *Geology* 40:695–698
- Klein FW (1982) Patterns of historical eruptions at Hawaiian volcanoes. *J Volcanol Geotherm Res* 12:1–35
- Klein FW (1984) Eruption forecasting at Kilauea Volcano, Hawaii. *J Geophys Res* 89:3059–3073
- Le Corvec N, Bebbington MS, Lindsay JM, McGee LE (2013) Age, distance, and geochemical evolution within a monogenetic volcanic field: analyzing patterns in the Auckland volcanic field eruption sequence. *Geochem Geophys Geosyst* 14:3648–3665
- Linde AT, Sacks IS (1998) Triggering of volcanic eruptions. *Nature* 395:888–890
- Lockwood JP (1995) Mauna Loa eruptive history – the preliminary radiocarbon record, Hawai‘i. In: Rhodes JM, Lockwood JP (eds) *Mauna Loa revealed: structure, composition, history, and hazards*, vol 92, *Geophysical monograph*. American Geophysical Union, Washington, DC, pp 81–94
- Luhr JF, Carmichael ISE (1990) Petrological monitoring of cyclical eruptive activity at Volcan Colima, Mexico. *J Volcanol Geotherm Res* 42:235–260
- Luongo G, Mazzarella A (2001) Is the state of the Vesuvius Volcano turbulent? *Mineral Petrol* 73:39–45
- Lutz TM (1986) An analysis of the orientation of large-scale crustal structures: a statistical approach based on areal distributions of pointlike features. *J Geophys Res* 91:421–434
- Lutz TM, Gutmann JT (1995) An improved method for determining and characterizing alignments of pointlike features and its implications for the Pinacate volcanic field, Sonora, Mexico. *J Geophys Res* 100: 17659–17670
- Magill CR, McAneney KJ, Smith IEM (2005) Probabilistic assessment of vent locations for the next Auckland volcanic field event. *Math Geol* 37:227–242
- Manga M, Brodsky E (2006) Seismic triggering of eruptions in the far field: volcanoes and geysers. *Ann Rev Earth Planet Sci* 34:263–291
- Marti J, Felpeto A (2010) Methodology for the computation of volcanic susceptibility. An example for mafic and felsic eruptions on Tenerife (Canary Islands). *J Volcanol Geotherm Res* 195:69–77
- Martin AJ, Umeda K, Connor CB, Weller JN, Zhao D, Takahashi M (2004) Modeling longterm volcanic hazards through Bayesian inference: an example from the Tohoku volcanic arc, Japan. *J Geophys Res* 109, B10208
- Martin DP, Rose WI (1981) Behavioral patterns of Fuego volcano, Guatemala. *J Volcanol Geotherm Res* 10:67–81
- Marzocchi W (1996) Chaos and stochasticity in volcanic eruptions the case of Mount Etna and Vesuvius. *J Volcanol Geotherm Res* 70:205–212
- Marzocchi W (2002) Remote seismic influence on large explosive eruptions. *J Geophys Res* 107:2018. doi:10.1029/2001JB000307
- Marzocchi W, Zaccarelli L (2006) A quantitative model for the time-size distribution of eruptions. *J Geophys Res* 111, B04204
- Marzocchi W, Scandone R, Mulargia F (1993) The tectonic setting of Mount Vesuvius and the correlation between its eruptions and the earthquakes of the southern Apennines. *J Volcanol Geotherm Res* 58:27–41
- Marzocchi W, Zaccarelli L, Boschi E (2004a) Phenomenological evidence in favor of a remote seismic coupling for large volcanic eruptions. *Geophys Res Lett* 31, L04601
- Marzocchi W, Sandri L, Gasparini P, Newhall C, Boschi E (2004b) Quantifying probabilities of volcanic events: the example of volcanic hazard at Mount Vesuvius. *J Geophys Res* 109, B11201
- Marzocchi W, Sandri L, Selva J (2010) BET VH: a probabilistic tool for long-term volcanic hazard assessment. *Bull Volcanol* 72:705–716
- Mason BG, Pyle DM, Dade WB, Jupp T (2004) Seasonality of volcanic eruptions. *J Geophys Res* 109, B04206
- Mauk FJ, Johnston MJS (1973) On the triggering of volcanic eruptions by earth tides. *J Geophys Res* 78: 3356–3362
- Mazzarini F, Ferrari L, Isola I (2010) Self-similar clustering of cinder cones and crust thickness in the Michoacan-Guanajuato and Sierra de Chichinautzin volcanic fields, Trans-Mexican Volcanic Belt. *Tectonophysics* 486:55–64
- Medina Martinez F (1983) Analysis of the eruptive history of the Volcan de Colima, Mexico (1560–1980). *Geof Int* 22:157–178
- Mendoza-Rosas AT, De La Cruz-Reyna S (2008) A statistical method linking geological and historical eruption time series for volcanic hazard estimations: applications to active polygenetic volcanoes. *J Volcanol Geotherm Res* 176:277–290
- Mendoza-Rosas AT, De La Cruz-Reyna S (2009) A mixture of exponentials distribution for a simple and precise assessment of the volcanic hazard. *Nat Haz Earth Syst Sci* 9:425–431

- Mulargia F (1992) Time association between series of geophysical events. *Phys Earth Planet Int* 71:147–153
- Mulargia F, Tinti S, Boschi E (1985) A statistical analysis of flank eruptions on Etna volcano. *J Volcanol Geotherm Res* 23:263–272
- Mulargia F, Gasperini P, Tinti S (1987) Identifying regimes in eruptive activity: an application to Etna volcano. *J Volcanol Geotherm Res* 34:89–106
- Mulargia F, Marzocchi W, Gasperini P (1992) Statistical identification of physical patterns which accompany eruptive activity on Mount Etna, Sicily. *J Volcanol Geotherm Res* 53:289–296
- Newhall CG, Self S (1982) The volcanic explosivity index (VEI): an estimate of explosive magnitude for historical volcanism. *J Geophys Res* 87:1231–1238
- Nishi Y, Inoue M, Tnaka T, Murai M (2001) Analysis of time sequences of explosive volcanic eruptions of Sakurajima. *J Phys Soc Jpn* 70:1422–1428
- Nostro C, Stein RS, Cocco M, Belardinelli ME, Marzocchi W (1998) Two-way coupling between Vesuvius eruptions and southern Apennine earthquakes, Italy, by elastic stress transfer. *J Geophys Res* 103:24487–24504
- Ogata Y (1988) Statistical models for earthquake occurrences and residual analysis for point processes. *J Am Stat Assoc* 83:9–27
- Orsi G, Di Vito MA, Selva J, Marzocchi W (2011) Long-term forecast of eruption style and size at Campi Flegrei caldera (Italy). *Earth Planet Sci Lett* 287:265–276
- Passarelli L, Sandri L, Bonazzi A, Marzocchi W (2010a) Bayesian Hierarchical Time Predictable Model for eruption occurrence: an application to Kilauea Volcano. *Geophys J Int* 181:1525–1538
- Passarelli L, Sanso B, Sandri L, Marzocchi W (2010b) Testing forecasts of a new Bayesian time-predictable model of eruption occurrence. *J Volcanol Geotherm Res* 198:57–75
- Pyle DM (1998) Forecasting sizes and repose times of future extreme volcanic events. *Geology* 26:367–370
- Reyment RA (1969) Statistical analysis of some volcanologic data regarded as series of point events. *Pure Appl Geophys* 74:57–77
- Rodado A, Bebbington M, Noble A, Cronin S, Jolly G (2011) On selection of analogue volcanoes. *Math Geosci* 43:505–519
- Salvi F, Scandone R, Palma C (2006) Statistical analysis of the historical activity of Mount Etna, aimed at the evaluation of volcanic hazard. *J Volcanol Geotherm Res* 154:159–168
- Sandri L, Marzocchi W, Gasperini P (2005) Some insights on the occurrence of recent volcanic eruptions of Mount Etna volcano (Sicily, Italy). *Geophys J Int* 163:1203–1218
- Santacroce R (1983) A general model for the behaviour of the Somma-Vesuvius volcanic complex. *J Volcanol Geotherm Res* 17:237–248
- Scandone R, Giacomelli L, Gasparini P (1993a) Mount Vesuvius: 2000 years of volcanological observations. *J Volcanol Geotherm Res* 58:5–25
- Scandone R, Argane G, Galdi F (1993b) The evaluation of volcanic risk in the Vesuvian area. *J Volcanol Geotherm Res* 58:263–271
- Scandone R, Giacomelli L, Speranza FF (2008) Persistent activity and violent strombolian eruptions at Vesuvius between 1631 and 1944. *J Volcanol Geotherm Res* 170:167–180
- Selva J, Costa A, Marzocchi W, Sandri L (2010) BET VH: Exploring the influence of natural uncertainties on long-term hazard from tephra fallout at Campi Flegrei (Italy). *Bull Volcanol* 72:717–733
- Selva J, Orsi G, Di Vito MA, Marzocchi W, Sandri L (2012) Probability hazard map for future vent opening at the Campi Flegrei caldera, Italy. *Bull Volcanol* 74:497–510
- Settle M, McGetchin TR (1980) Statistical analysis of persistent explosive activity at Stromboli, 1971: implications for eruption prediction. *J Volcanol Geotherm Res* 8:45–58
- Sharp ADL, Lombardo G, David PM (1981) Correlation between eruptions of Mount Etna, Sicily, and regional earthquakes as seen in historical records from AD 1582. *Geophys J Roy Astron Soc* 65:507–523
- Sheridan MF (1992) A Monte Carlo technique to estimate the probability of volcanic dikes. In: High-level radioactive waste management: proceedings of the third annual international conference, Las Vegas, 12–16 Apr 1992, American Nuclear Society, La Grange Park, pp 2033–2038
- Shimazaki K, Nakata T (1980) Time-predictable recurrence model for large earthquakes. *Geophys Res Lett* 7:279–282
- Siebert L, Simkin T (2002) Volcanoes of the world: an illustrated catalog of Holocene volcanoes and their eruptions. Smithsonian Institution, Global Volcanism Program Digital Information Series, GVP-4. <http://www.volcano.si.edu/index.cfm>
- Simkin T (1993) Terrestrial volcanism in space and time. *Ann Rev Earth Planet Sci* 21:427–452
- Simkin T (1994) Distant effects of volcanism – how big and how often? *Science* 264:913–914
- Smethurst L, James MR, Pinkerton H, Tawn JA (2009) A statistical analysis of eruptive activity on Mount Etna, Sicily. *Geophys J Int* 179:655–666
- Smith EI, Keenan DL, Plank T (2002) Episodic volcanism and hot mantle: implications for volcanic hazard studies at the proposed nuclear waste repository at Yucca Mountain, Nevada. *GSA Today* 12 (April):4–9
- Sobradelo R, Geyer A, Marti J (2010) Statistical data analysis of the CCDB (Collapse Caldera Database): insights on the formation of caldera systems. *J Volcanol Geotherm Res* 198:241–252
- Sobradelo R, Marti J (2010) Bayesian event tree for long-term volcanic hazard assessment: application to Teide-Pico Viejo stratovolcanoes, Tenerife, Canary Islands. *J Geophys Res* 115, B05206
- Solow AR (1993) Estimating record inclusion probability. *Am Stat* 47:206–208
- Solow AR (2001) An empirical Bayes analysis of volcanic eruptions. *Math Geol* 33:95–102
- Sornette A, Dubois J, Cheminee JL, Sornette D (1991) Are sequences of volcanic eruptions deterministically chaotic? *J Geophys Res* 96:11931–11945
- Stothers RB (1989) Seasonal variations of volcanic eruption frequencies. *Geophys Res Lett* 16:453–455

- Takada A (1997) Cyclic flank-vent and central-vent eruption patterns. *Bull Volcanol* 58:539–556
- Tanguy J-C, Condomines M, Le Goff M, Chillemi V, La Delfa S, Pantane G (2007) Mount Etna eruptions of the last 2,750 years: revised chronology and location through archeomagnetic and ^{226}Ra - ^{230}Th dating. *Bull Volcanol* 70:55–83
- Telesca L, Lapenna V (2005) Identifying features in time-occurrence sequences of volcanic eruptions. *Environmetrics* 16:181–190
- Telesca L, Cuomo V, Lapenna V, Macchiato M (2002) Time-clustering analysis of volcanic occurrence sequences. *Phys Earth Planet Int* 131:47–62
- Thorlaksson JE (1967) A probability model of volcanoes and the probability of eruptions of Hekla and Katla. *Bull Volcanol* 31:97–106
- Turner M, Cronin S, Bebbington M, Platz T (2008a) Developing a probabilistic eruption forecast for dormant volcanoes; a case study from Mt Taranaki, New Zealand. *Bull Volcanol* 70:507–515
- Turner M, Cronin S, Smith I, Bebbington M, Stewart RB (2008b) Using titanomagnetite textures to elucidate volcanic eruption histories. *Geology* 36:31–34
- Turner M, Bebbington M, Cronin S, Stewart R (2009) Merging eruption datasets: building an integrated Holocene eruptive record of Mt Taranaki. *Bull Volcanol* 71:903–918
- Turner MB, Cronin SJ, Bebbington MS, Smith IEM (2011) Andesitic tephrochronology: construction of a pyroclastic eruption record for Mt Taranaki, New Zealand. *Quat Int* 246:364–373
- Valentine GA, Perry FV (2007) Tectonically controlled, time-predictable basaltic volcanism from a lithospheric mantle source (central Basin and Range Province, USA). *Earth Planet Sci Lett* 261:201–216
- Vere-Jones D (1992) Statistical methods for the description and display of earthquake catalogs. In: Walden AT, Guttorp P (eds) *Statistics in the environmental and earth sciences*. Edward Arnold, London, pp 220–246
- Vere-Jones D (1998) Probabilities and information gain for earthquake forecasting. *Comput Seismol* 30:248–263
- Voight B (1988) A method for prediction of volcanic eruptions. *Nature* 332:125–130
- Wadge G (1982) Steady state volcanism: evidence from eruption histories of polygenetic volcanoes. *J Geophys Res* 87:4035–4049
- Wadge G, Burt L (2011) Stress field control of eruption dynamics at a rift volcano: Nyamuragira, D.R. Congo. *J Volcanol Geotherm Res* 207:1–15
- Wadge G, Cross A (1988) Quantitative methods for detecting aligned points: an application to the volcanic vents of the Michoacan-Guanajuato volcanic field, Mexico. *Geology* 16:815–818
- Wadge G, Guest JE (1981) Steady-state magma discharge at Etna 1971–1981. *Nature* 294:548–550
- Wadge G, Walker WPL, Guest JE (1975) The output of Etna volcano. *Nature* 255:385–387
- Wang T, Bebbington M (2012) Estimating the likelihood of an eruption from a volcano with missing onsets in its record. *J Volcanol Geotherm Res* 243–244:14–23
- Wang T, Bebbington M (2013) Robust estimation for the Weibull process applied to eruption records. *Math Geosci* 45:851–872
- Watt SFL, Mather TA, Pyle DM (2007) Vulcanian explosion cycles: patterns and predictability. *Geology* 35: 839–842
- Watt SFL, Pyle DM, Mather TA (2009) The influence of great earthquakes on volcanic eruption rate along the Chilean subduction zone. *Earth Planet Sci Lett* 277: 399–407
- Weller JN (2004) Bayesian inference in forecasting volcanic hazards: an example from Armenia. Unpublished MS thesis, University of South Florida
- Wickman FE (1966a) Repose-period patterns of volcanoes. I. Volcanic eruptions regarded as random phenomena. *Arch Mineral Geol* 4:291–367
- Wickman FE (1966b) Repose-period patterns of volcanoes. V. General discussion and a tentative stochastic model. *Arch Mineral Geol* 4:351–367
- Wickman FE (1976) Markov models of repose-period patterns of volcanoes. In: Merriam DF (ed) *Random processes in geology*. Springer, New York, pp 135–161
- Young PC (2006) New approaches to volcanic time-series analysis. In: Mader HM, Coles SG, Connor CB, Connor LJ (eds) *Statistics in volcanology*. Geological Society, London, pp 143–160
- Zernack AV, Cronin SJ, Bebbington MS, Price RC, Smith IEM, Stewart RB, Procter JN (2012) Forecasting catastrophic stratovolcano collapse, a model based on Mt. Taranaki, New Zealand. *Geology* 40:983–986

Books and Reviews

- Bebbington MS (2013) Models for temporal volcanic hazard. *Stat Volcanol* 1:1–24, doi: <http://dx.doi.org/10.5038/2163-338X.1.1>
- Connor CB, Chapman NA, Connor LJ (eds) (2009) *Volcanic and tectonic hazard assessment for nuclear facilities*. Cambridge University Press, Cambridge
- Guttorp P (1995) *Stochastic modeling of scientific data*. Chapman and Hall, London
- Hill DP, Pollitz FP, Newhall C (2002) Earthquake-volcano interactions. *Phys Today* 55:41–47
- Lindsay JK (2004) *Statistical analysis of stochastic processes in time*. Cambridge University Press, Cambridge
- Marzocchi W, Bebbington MS (2012) Probabilistic eruption forecasting at short and long time scales. *Bull Volcanol* 74:1777–1805
- Mader HM, Coles SG, Connor CB, Connor LJ (eds) (2006) *Statistics in volcanology*. Geological Society, London

Volcanic Hazards and Early Warning

Robert I. Tilling
Volcano Science Center, US Geological Survey,
Menlo Park, USA

Article Outline

Glossary
Definition of the Subject
Introduction
Scope of Problem: Challenge for Emergency-
Management Authorities
Variability in Possible Outcomes of Volcano
Unrest
Some Recent Examples of Actual Outcomes of
Volcano Unrest
Challenges in Achieving Refined Predictive
Capability
Future Directions
Bibliography

Glossary

“Baseline”-monitoring data Volcano-monitoring data acquired for a volcanic system documenting its range of variation during its “normal” behavior prior to onset of volcano unrest and (or) eruptions. The longer the time span covered by “baseline” monitoring, the more diagnostic the dataset for identifying any significant departures from normal behavior in anticipating the possible outcomes of escalating volcano unrest.

“Decision window” The period of time – typically weeks to months but can be longer – during a developing volcanic crisis after the onset or escalation of volcano unrest. During this time, emergency managers are under high-stress political and public pressure to make

decisions regarding mitigative actions, including possible evacuation of populations at risk.

Factual statement Following the recommended definition of Swanson et al. (1985), it is the description of the current status and conditions of a volcano but does not anticipate future events.

Forecast As defined by Swanson et al. (1985), it is a comparatively imprecise statement of the time, place, and nature of expected activity; forecasts of eruptions and earthquakes sometimes are probabilistic (e.g., Turner et al. 2008; WG99, Working Group on California Earthquake Probabilities 1999). A forecast usually covers a longer time period than that for a prediction.

Magma intrusion The subsurface movement or injection of *magma* (molten rock containing associated crystals and gases) from one part of a volcanic system into another. Typically, an intrusion involves transport of magma from a central zone of storage (i.e., magma “reservoir”) into peripheral, structurally weaker zones (e.g., faults or rifts). Some intrusions culminate in surface eruptive outbreaks, others do not.

Prediction As defined by Swanson et al. (1985), it is a comparatively precise statement of the time, place, and, ideally, the nature and size of impending activity. Forecasts and predictions can be either *long term* (typically years to decades or longer) or *short term* (typically hours to months).

Volcanic crisis An unstable situation or time of heightened public concern when the level of volcano unrest exceeds its “baseline” level, thereby increasing the prospects of possible eruption at some future but indeterminate time. In general, during a crisis, emergency managers face a relatively narrow “decision window” in which to take timely mitigative actions to ensure public safety.

Volcano hazards Potentially damaging volcano-related processes and products that occur

during or following eruptions (see (Sigurdsson et al. 2000; Tilling 2002) for overviews). In quantitative hazards assessments, the probability of a given geographic area or region being affected by potentially destructive phenomena within a given period of time.

Volcano monitoring The systematic collection, analysis, and interpretation of visual observations and instrumental measurements of changes at volcanoes before, during, and after the onset of volcano unrest and (or) eruptive activity.

Volcano risk Probability of harmful consequences – individual or societal – or expected losses (deaths, injuries, property, livelihoods, economic activity disrupted or environment damaged) resulting from interactions between volcano hazards, human development, and vulnerable conditions. Though definitions vary in detail, risk is conventionally expressed by the general relation: $risk = hazards \times vulnerability$ (ISDR 2004).

Volcano status The current activity or mode of behavior of a volcano. Status is commonly described as follows: *active* (having one or more eruptions during recorded history), *dormant* (currently inactive but having potential for renewed eruptive activity), and *extinct* (dormant for long time and not expected to erupt again). These terms, while commonly used in the scientific literature, are imperfect and are undergoing serious reexamination within the global volcanological community.

Volcano unrest Visual and (or) measurable physical and (or) chemical changes – surface or subsurface – in the status of the volcano, relative to its “normal” historical behavior; duration of unrest can vary from hours to decades. The initiation or escalation of volcano unrest, regardless of duration, may or may not culminate in eruption.

Vulnerability The conditions determined by physical, social, economic, and environmental factors or processes, which increase the susceptibility of a community to the adverse impacts of hazards (ISDR 2004).

Warning An official message issued by government authorities, usually given to a specific

community or communities when a direct response to a volcanic threat is required. To be useful, warnings must be credible and effectively communicated – in clear, easily understandable language – with sufficient lead time, ideally, well before the volcano unrest escalates into a volcanic crisis.

Definition of the Subject

The hazards and risks posed by volcanic eruptions are increasing inexorably with time. This trend is the direct result of continuing exponential growth in global population and progressive encroachment of human settlement and economic development into hazardous volcanic areas. One obvious strategy in reducing volcano risk is the total abandonment of hazardous volcanic regions for human habitation. Clearly, this is utterly unrealistic; many hazardous volcanoes are located in already densely populated areas, for most of which land-use patterns have been fixed by history, culture, and tradition for centuries or millennia. Moreover, people also are exposed to potential volcano hazards by simply being passengers aboard commercial airliners flying over volcanic regions and possibly encountering – while in flight – a drifting volcanic ash cloud from a powerful explosive eruption (Casadevall 1994; Guffanti et al. 2010; OFCM 2004). Fortunately, to date no lives have been lost from volcanic ash-aircraft encounters, but some of the most severe incidents have caused major economic losses and disruptions to world air traffic. For example, in mid-April 2010, the high eruption cloud from the Eyjafjallajökull Volcano (Iceland) prompted aviation officials to close – as a precautionary measure – many airports in Europe; many Europe-bound flights from the United States were also cancelled (Alexander 2013). The airport closures and flight cancellations resulted in the biggest and costliest disruption in commercial aviation since that caused by the 11 September 2001, terrorist attacks on the United States (Alexander 2013 and references therein).

The only viable option in reducing volcano risk is the timely issuance of early warning of possible

impending eruptions, allowing emergency-management officials to take mitigative actions. It is then imperative for volcanologists to effectively communicate the best-possible hazards information to emergency-management authorities and to convince them to implement timely mitigative countermeasures, including evacuation, if necessary, of people at risk. However, volcanologists face daunting challenges in providing reliable, precise early warnings that government officials demand and need to make informed decisions to ensure public safety.

Introduction

About 1500 volcanoes have erupted one or more times during the Holocene (i.e., past 10,000 years); since A.D. 1600, volcanic disasters have killed about 300,000 people and resulted in property damage and economic loss exceeding hundreds of millions of dollars (Tilling 2005, 2008). During recent centuries, on average 50 to 70 volcanoes worldwide are in eruption each year – about half representing activity continuing from the previous year and the other half new eruptive outbreaks (Siebert et al. 2010). Most of these erupting volcanoes are located in developing countries, which often lack sufficient scientific and economic resources to conduct adequate volcano-monitoring studies. Moreover, in any given year, many more volcanoes exhibit observable and (or) measurable anomalous behavior (*volcano unrest*) that does not lead to eruption.

For scientists and emergency-management authorities alike, the critical issues in providing early warning of possible eruption, or of possible escalation of an ongoing but weak eruption, involve basic questions such as (1) will the initiation of unrest at a currently dormant volcano culminate in eruption?; (2) if so, what would be its time of onset, eruptive mode (explosive, non-explosive, etc.) and duration?; (3) at an already restless or weakly erupting volcano, what are the prospects of escalation to major activity?; and (4) what would be the nature and severity of the associated hazards and which sectors of the volcano are the most vulnerable?

Systematic geoscience studies of a volcano's geologic history and past behavior can provide answers to some of the above-listed and related questions relevant to *long-term* forecasts of a volcano's present and possible future activity. In general, the fundamental premise applied in making long-term forecasts is that the volcano's past and present behavior provides the best clues to its future behavior. However, data from volcano monitoring – showing increased activity relative to pre-crisis or “background” levels – constitute the only scientific basis for *short-term* forecasts or predictions of a possible future eruption or a significant change during an ongoing eruption. To respond effectively to a developing volcanic crisis, timely early warnings are absolutely essential, and these warnings must be reliable and precise. Improvement in the reliability and precision of warnings can be achieved only by a greatly improved capability for eruption prediction, which in turn depends on the quantity and quality of volcano-monitoring data and the diagnostic interpretation of such information (Tilling 1995, 2003, 2008).

At the outset, it must be emphasized that, in this article, the term *prediction* is used in the strict sense (see Glossary) proposed by Swanson et al. (1985); thus, for a forecast or prediction to be genuine, it must be included in an official warning statement made publicly in advance of the forecasted or predicted event. While retrospective forecasts and predictions serve valid scientific purposes in testing or verifying theories or models, they are of little or no use to civil authorities during the actual management of a volcanic crisis. It is beyond the scope of this brief paper to consider in detail volcano-monitoring studies and their applications to eruption prediction. Suffice it to say that instrumental volcano monitoring is a multifaceted, multidisciplinary endeavor that involves the ground- and (or) satellite-based measurement of seismicity, ground deformation, gas and fluid chemistry, temperature variations, secular changes in microgravity and geomagnetism, etc. For more information about volcano-monitoring approaches, techniques, and instrumentation, interested readers are referred to numerous recent comprehensive reviews and

references cited therein (e.g., Chouet 2004; Dzurisin 2006; McNutt 1996, 2000a, b; Tilling 1995; ► “[Volcanoes, Non-linear Processes in](#)”). Instead, this entry highlights the range in possible outcomes of volcano unrest and reviews some recent examples of the actual outcomes documented for several well-monitored volcanoes. Some implications from the observations discussed in this entry are considered in the final two parts of this entry (sections “[Challenges in Achieving Refined Predictive Capability](#)” and “[Future Directions](#)”).

Scope of Problem: Challenge for Emergency-Management Authorities

Because earthquake and eruption dynamics intrinsically involve nonlinear processes (the focus of this volume), the attainment of a reliable capability to predict eruptions (especially explosive events) still eludes volcanologists. This dilemma persists despite notable advances in volcanology and the availability in recent decades of increasingly sophisticated models, stochastic and deterministic. Such models, which commonly involve pattern recognition methodologies, are developed from theoretical or statistical analysis of volcano-monitoring data, eruption-occurrence patterns, or other time-space attributes for a volcanic system (e.g., Klein 1982, 1984; Marzocchi et al. 2004; Murray and Ramirez Ruiz 2002; Sandri et al. 2004; Sparks 2003; Voight 1988; Voight and Cornelius 1991; ► “[Volcanic Eruptions: Stochastic Models of Occurrence Patterns](#)”). The non-attainment to date of predictive capability reflects the reality that most volcanic systems are highly complex and that the relevant datasets for the vast majority of volcanoes are too inadequate and (or) incomplete to enable reliable and precise predictions.

Given the fact that the emerging science of eruption prediction is still nascent, scientists and emergency-management officials are at a substantial disadvantage when a long-dormant or weakly active volcano begins to exhibit unrest above its historical (i.e., “normal”) levels. A principal concern is that the heightened unrest, should it persist

or escalate, could culminate in renewed or greatly increased, possibly hazardous activity. Because of public safety and socioeconomic issues and consequences, emergency managers in particular must have a definitive answer to the question: What is the most likely outcome of the escalating unrest? The answer(s) to this critical question then shape(s) possible decisions they must make – within a narrow “decision window” (typically weeks to months), generally under confusing, stressful conditions – in managing the growing volcanic crisis.

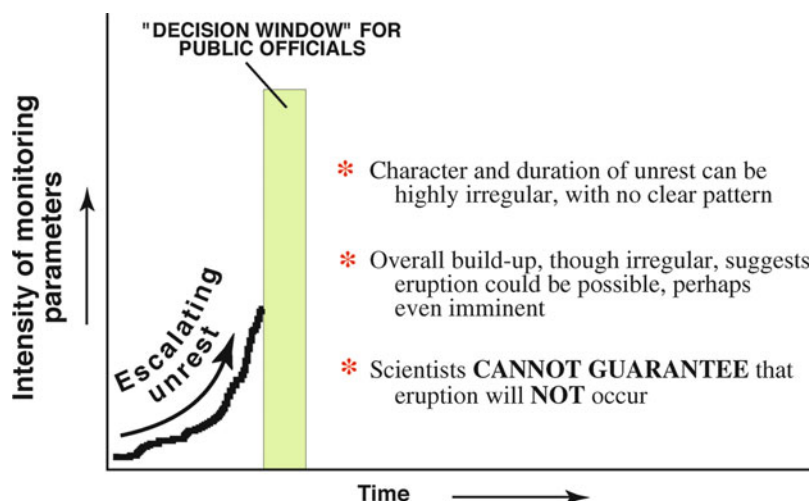
Figure 1 depicts a common scenario during a volcanic crisis and lists some major factors that complicate the determination of the most likely outcome of the unrest. Unfortunately, despite advances in volcanology and volcano-monitoring techniques, scientists are unable to give emergency-management authorities the specific information they require. The greatest frustration for the authorities is that, given the current state of the art of volcanology, scientists cannot guarantee that an eruption will *not* occur. In the section below, I review some common possible outcomes of volcano unrest and, in section “[Some Recent Examples of Actual Outcomes of Volcano Unrest](#),” some actual outcomes of recent episodes of volcano unrest.

Variability in Possible Outcomes of Volcano Unrest

Experience worldwide for historical eruptions demonstrates a great range in possible outcomes of volcano unrest. The three types of outcomes discussed below – for which adequate eruptive history and volcano-monitoring data exist – probably represent the commonest possibilities. Given the nonlinear nature of eruptive phenomena, however, other scenarios can be envisaged.

Culmination in Major Eruption or Return to Dormancy Following Short Duration of Unrest

Perhaps the simplest scenario – and certainly the most easily manageable for public officials – is when a volcano reawakens from a long dormancy (i.e., greater than a century) and begins to exhibit



Volcanic Hazards and Early Warning, Fig. 1 A common scenario confronting emergency-management authorities when unrest initiates or escalates at a volcano, as documented by increasing intensity of monitoring parameters (seismicity, ground deformation, gas emission, thermal, microgravity, etc.) during a volcanic crisis. The “decision window,” though shown schematically,

represents a very real and usually short time frame during which emergency managers must decide and implement – by legal requirement, political and (or) citizen pressure, or moral conscience – strategies and actions to mitigate the potential risk from volcano hazards (Figures 1 and 2 modified from unpublished conceptual sketches of C. Dan Miller, US Geological Survey)

significant unrest. Such unrest then persists at a high level for a relatively short duration (several weeks or months) before culminating in a major eruption (Fig. 2a). With adequate volcano-monitoring studies, preferably in real time or near-real time, conducted during the precursory buildup, scientists may be able to make a short-term forecast or prediction of the impending eruption. The emergency-management authorities then could have the needed guidance during the “decision window” in issuing early warnings and implementing mitigative countermeasures. Alternatively, diagnostic analysis of the volcano-monitoring data, combined with detailed knowledge of eruptive history, could lead the scientists to conclude that an eruption is not likely. In this situation, emergency managers also receive the required scientific input they need to make informed decisions to manage the volcanic crisis, including the declaration of the end of the crisis and the cancellation of any warnings in effect. In either case, scientists need sufficient information to try to determine whether an imminent eruption or a return of the volcano to dormancy is the most likely outcome. However, it must be emphasized that scientists in assessing possible outcomes can

reach incorrect conclusions, even using abundant scientific data and constraints from quantitative models.

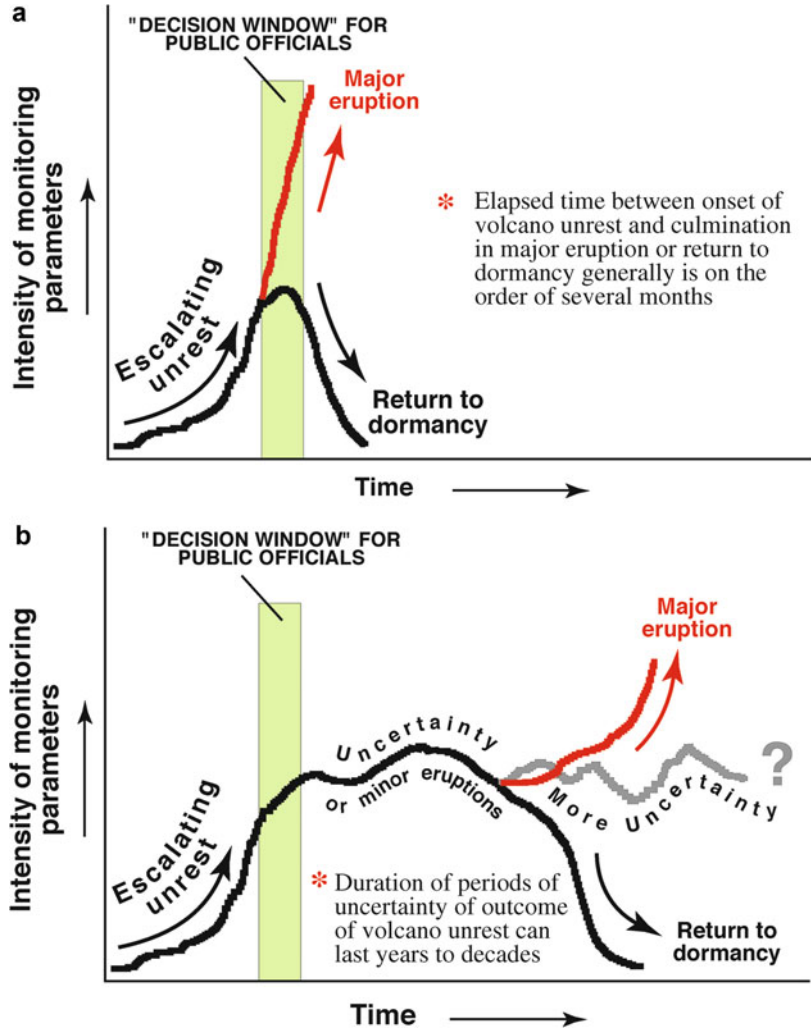
Culmination in Major Eruption or Return to Dormancy Following a Long Lull in Unrest

A scenario fraught with greater uncertainty is one that, after an initial short period (weeks to months) of sustained and heightened volcano unrest, the volcano neither erupts nor quickly returns to quiescence. Instead, the volcanic system exhibits a decreased level of activity characterized by fluctuating but continuing low-level unrest and (or) weak, nonhazardous eruptions that may persist for years or decades, before possibly exhibiting another episode of heightened unrest. Under such circumstances, scientists and public officials confront anew the crisis conditions and challenges inherent in attempting to anticipate the outcome of the renewed unrest.

This scenario is similar to that previously discussed above (section “[Culmination in Major Eruption or Return to Dormancy Following Short Duration of Unrest](#)”) but differs by the

Volcanic Hazards and Early Warning, Fig. 2

(a) Two common possible outcomes of volcano unrest that are arguably the most tractable for public officials managing a volcanic crisis (see text). (b) Some other possible outcomes of volcano unrest that pose much greater challenges for scientists and public officials alike during a volcanic crisis (see text)



complication posed by another possible outcome in addition to those of an imminent eruption or a return to dormancy, namely, the possibility that the volcano could revert to a mode of continuing uncertainty involving only low-level activity or cessation of unrest for an indeterminate period of time (Fig. 2b). The prolongation of the periods of uncertainty has some unfortunate consequences, including (1) the potential for scientists and public officials to lose credibility by incorrectly anticipating the outcomes during the pulses of heightened unrest and by issuing warnings ("false alarms") of events that fail to materialize (see section "The Dilemma of 'False Alarms'")

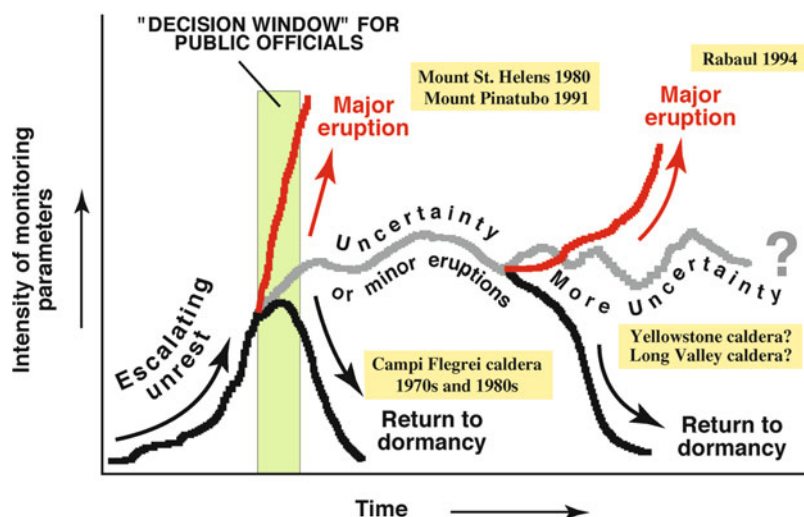
and (2) increased difficulty in maintaining public interest in, and awareness of, potential hazards posed by some eventual volcanic crisis that actually culminates in eruption.

Ongoing Unrest over Long Periods with No Clear Indication of Possible Future Activity

Some long-dormant volcanic systems undergo unrest that can persist for many years to decades, fluctuating in intensity, but not showing any long-term definitive trend suggestive of possible renewed eruptive activity (Fig. 2b). It is extremely difficult, perhaps impossible, to anticipate possible outcomes of such long-duration, ongoing

Volcanic Hazards and Early Warning,

Fig. 3 Some actual outcomes of recent episodes of volcano unrest at selected volcanoes. Even though these well-studied volcanoes are closely monitored in real time or near-real time by instrumental arrays, it was or is not possible to determine the outcome of the unrest at most of them (see text for discussion)



unrest. The reasons for this problem are simply that the eruptive histories and dynamics of long-dormant systems are poorly understood and (or) their eruption frequencies are longer than the time span of the available pre-eruption *baseline*-monitoring data. Some examples of such modes of volcano unrest include Three Sisters volcanic center, Oregon, USA (Wicks et al. 2002); Long Valley Caldera, California, USA (Battaglia et al. 2003; Hill 2006); Yellowstone Caldera, Wyoming-Montana, USA (Wicks et al. 2006); and Campi Flegrei Caldera, Italy (Berrino et al. 1984; Troise et al. 2007).

Some Recent Examples of Actual Outcomes of Volcano Unrest

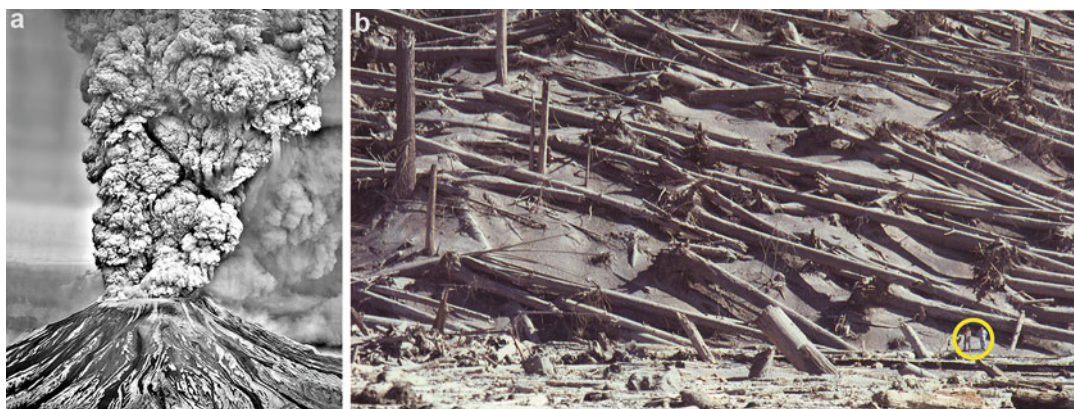
The previous discussion of possible hypothetical outcomes of volcano unrest (section “[Variability in Possible Outcomes of Volcano Unrest](#)”) has its basis in several recent examples of actual volcano unrest, as documented by detailed investigations. These examples are reviewed herein (Fig. 3).

Mount St. Helens (USA) and Mount Pinatubo (Philippines)

In 1975, on the basis of many years of detailed stratigraphic and dating studies, together with an analysis of repose intervals, scientists of the US Geological Survey (USGS) made a long-term

forecast of possible renewed eruptive activity at Mount St. Helens Volcano, Washington State, which had been dormant since 1857. This forecast, albeit qualitative, was remarkably prescient, stating: “...an eruption is likely within the next 100 years, possibly before the end of this century” (see p. 441 in Crandell et al. 1975). Five years after the forecast, phreatic eruptions began at Mount St. Helens in late March 1980 following a week of precursory seismicity. Sustained seismicity, accompanied by large-scale ground deformation of the volcano’s north flank, ultimately produced a cataclysmic magmatic eruption on 18 May 1980 (Fig. 4). This catastrophic event caused the worst volcanic disaster in the history of the United States (see (Lipman and Mullineaux 1981) and chapters therein for detailed summary; Tilling et al. (1990); (Tilling 2000)).

For Mount St. Helens, the volcano unrest, which was well monitored visually and instrumentally, lasted for only about 2 months before culminating in a major eruption (Fig. 3). Scientists monitoring Mount St. Helens were not able to make a precise prediction of the 18 May events (flank collapse, laterally directed blast, and vertical magmatic eruption plume). From analysis of the volcano-monitoring data and hazards assessments, however, they were able to give early warning and to convince emergency-management authorities to order evacuation and take other risk-reduction actions well within the “decision



Volcanic Hazards and Early Warning, Fig. 4 (a) 19-km-high eruption plume of the climactic magmatic eruption of Mount St. Helens Volcano, Washington State, USA, on 18 May 1980. This eruption occurred following about 2 months of volcano unrest that began after a

123-year quiet period (Photograph by Austin Post taken about noon on 18 May 1980). (b) Douglas fir trees flattened by the tremendous force of the 18 May lateral blast; two people (*circled*, lower right-hand corner) indicate scale (USGS Photograph by Lyn Topinka)

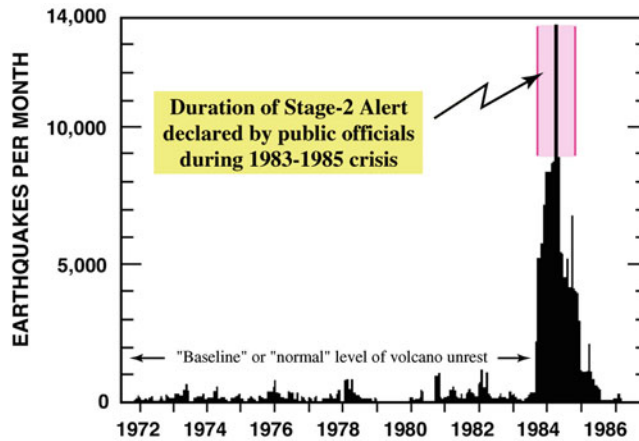
window” prior to the paroxysmal eruption. Had timely mitigative actions not been taken, fatalities from the 18 May eruption almost certainly would have been higher than 57. After 18 May, eruptive activity continued episodically and non-explosively through mid-1986, constructing a series of lava domes within the newly formed crater (Lipman and Mullineaux 1981; Tilling et al. 1990; Tilling 2000). Following a period of quiescence (1986–2003), Mount St. Helens reactivated with a series of nonexplosive, dome-building eruptions during 2004–2006 that posed no serious volcanic hazards (Sherrod et al. 2008).

Another illustrative example of a major eruptive activity following a brief duration of volcano unrest was the 1991 eruption of Mount Pinatubo, Luzon, Philippines (see Newhall and Punongbayan 1996, and chapters therein). Prior to 1991, Pinatubo had been quiescent for about 600 years with no recorded historical eruptions; it had been little studied and was not monitored. In March 1991, Pinatubo began to stir from its long sleep, as indicated by increased fumarolic activity and precursory earthquakes, some large enough to be felt by local inhabitants. Phreatic explosions began in early April, and scientists of the Philippine Institute of Volcanology and Seismology (PHIVOLCS) and the USGS Volcano Disaster Assistance Program (VDAP) quickly established a rudimentary

volcano-monitoring network (Fig. 5) and prepared a hazard zonation map, under difficult and stressful conditions of an escalating volcanic crisis. Systematic monitoring of seismicity and emission of sulfur dioxide (SO_2) became operational by early May. The first magmatic activity (dome intrusion) was observed on 7 June. On the basis of the analysis and interpretation of the monitoring data, combined with observation of increased frequency of high (>15 km) eruption plumes, the PHIVOLCS-USGS scientific team issued warnings and made a forecast on 12 June that a major eruption was imminent. The PHIVOLCS-USGS scientific team recommended to the Commandant of the US Clark Air Base and to the Philippine emergency-management authorities to order major evacuations of the air base and other population centers around the volcano. In all, more than 300,000 people were evacuated from hazardous zones. Three days later on 15 June, Pinatubo’s climactic eruption occurred. Even though this eruption was the largest in the world since the 1912 eruption of Novarupta-Katmai (Alaska, USA), the timely early warnings and evacuations of people and expensive but moveable property (e.g., aircraft, vehicles, equipment, computers, household belongings) by military and civilian officials saved many thousands of lives and averted hundreds of millions of dollars in potential economic loss.

Volcanic Hazards and Early Warning,

Fig. 5 Members of the PHILVOLCS-USGS scientific team unloading equipment to establish the volcano-monitoring network at Mount Pinatubo, Luzon, Philippines, in response to the onset of volcano unrest in March 1991 (see text). Photograph by Val Gempis, US Air Force Combat Camera



Volcanic Hazards and Early Warning,
Fig. 6 Occurrences of earthquakes at Rabaul Caldera, Papua New Guinea, during the period 1972–1986 (Modified from McKee et al. 1985). The dramatically increased seismicity during 1983–1985, together with significant ground deformation, constituted a serious volcanic

crisis and prompted public officials to declare a “Stage 2 Alert” (see text). The time span bracketed by this alert (*pink shading*) is somewhat comparable to the “decision window” schematically shown in Figs. 1, 2, and 3. However, the 1983–1985 unrest did not immediately culminate in eruption, which did occur 10 years later

Rabaul Caldera (Papua New Guinea)

The outcomes for the volcano unrest at Mount St. Helens (1980) and Pinatubo (1991) represent examples of simple (linear?) and relatively short progression (several months) from onset of precursory activity to major eruption. However, most other recent episodes of volcano unrest involve complex or uncertain outcomes; below I discuss one well-documented example of a complex outcome.

Rabaul caldera, Papua New Guinea, after being quiescent for several decades began to exhibit volcano unrest in the mid-1980s, as manifested by dramatically increased seismicity (Fig. 6) and accompanying caldera uplift of more than 1 m (McKee et al. 1985). This intense and rapidly escalating unrest prompted the scientists of the Rabaul Volcanological Observatory (RVO) to warn of possible imminent eruption. In response, emergency-management authorities

declared in October 1983 a “Stage 2 Alert” (Fig. 6), stating that the volcano could erupt within weeks. Accordingly, the civil authorities implemented a number of mitigative measures – including widening of roads to serve as evacuation routes, preparation of contingency plans, conduct of evacuation drills, etc. – to prepare for the expected soon-approaching eruption. However, soon after the issuance of the alert, the volcano unrest, after peaking in early 1985, began to diminish sharply and returned to normal levels within 6 months (Fig. 6). No eruption occurred, and the “Stage 2 Alert” was lifted in November 1984.

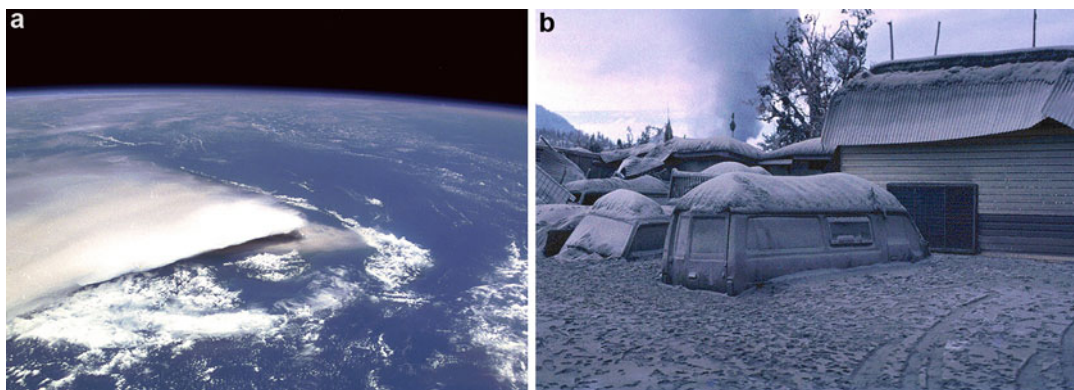
The abrupt cessation of volcano unrest and the nonoccurrence of the forecasted eruption were generally viewed as a needless, disruptive “false alarm.” As fallout from the unanticipated turn of events, the scientists of the RVO lost credibility with the emergency-response authorities, who in turn lost the confidence of the general population.

Ultimately however, the initially unsatisfactory outcome of the 1983–1985 unrest positively influenced the successful response to a later major eruption. After returning to, and remaining at, the normally low level of activity for about 10 years after the mid-1980s crisis, Rabaul erupted suddenly on 19 September 1994 following only 27 h of precursory seismicity (Blong and McKee 1995). This major eruption (Fig. 7) devastated nearly all of Rabaul City (Fig. 7b) but fortunately caused only five deaths. The much

lower-than-expected fatalities can in part be attributable to the fact that people largely “self-evacuated” without waiting for official advisories to do so. They apparently well remembered the lessons (e.g., enhanced awareness of volcano hazards, measures to take in case of eruption, evacuation drills) learned from the volcanic crisis 10 years earlier.

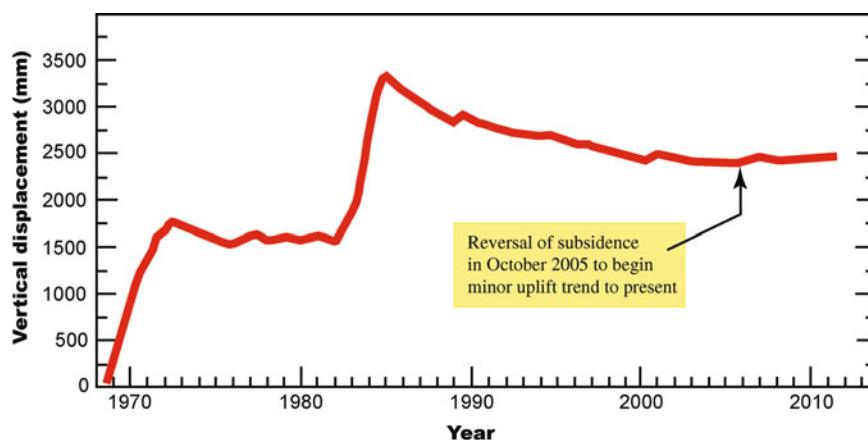
Ongoing Irregular Long-Duration Volcano Unrest but No Clear Indication of Possible Eruption

Recent volcano unrest at several calderas, which are well studied and systematically monitored in real time or near-real time by geophysical and geochemical networks, provide good examples of the difficulty in determining the most likely outcomes of unrest. Because of the abundance of historical observations and volcano-monitoring data, Campi Flegrei Caldera, Italy, located adjacent to Vesuvius Volcano in the Naples metropolitan area, is a particularly instructive example. During the early 1970s and again during 1983–1985, the caldera exhibited pronounced episodes of volcano unrest (Fig. 7) that generated scientific and public concern. Neither of these unrest episodes culminated in eruption, and each was followed by longer intervals of relatively low, fluctuating activity or net ground subsidence, hinting of an outcome whereby the restless volcano would return to dormancy.



Volcanic Hazards and Early Warning, Fig. 7 (a) View from space of the powerful explosive eruption of 19 September 1994 from Rabaul Caldera (Photograph courtesy of astronauts aboard the NASA Shuttle). (b)

Typical scene in Rabaul City showing the devastation caused by heavy ashfall from this eruption, which fortunately only caused few deaths (see text) (Photograph by Andy Lockhart, VDAP/USGS)



Volcanic Hazards and Early Warning, Fig. 8 Uplift and subsidence at Campi Flegrei Caldera, Italy, during the period 1969–2012 (Modified from Fig. 1b in Troise et al. 2007 and updated with data from Osservatorio 2014). Vertical displacements measured by precise leveling, together with associated horizontal displacements and

increased seismicity, indicate two episodes of caldera uplift, one during 1969–1973 and another during 1982–1985. Neither episode culminated in eruption (see text). Note that the reference datum of the current “baseline” level of activity is higher than that prior to 1982

From a recent study by Troise et al. (2007), however, the post-1985 subsidence trend reversed in late 2005, and the caldera since has risen by a few centimeters through 2011 (Fig. 8). Historical records, extending back in time for more than two millennia (see Fig. 1a in Troise et al. 2007), demonstrate a pattern of relatively short-duration uplifts (lasting years to decades) interspersed with long-lived gradual subsidence (lasting centuries). Of these several ups and downs of Campi Flegrei caldera over the past two millennia, only one episode of volcano unrest culminated in eruptive activity – the Monte Nuovo eruption in 1538 (De Vito et al. 1987). The historical behavior of Campi Flegrei emphasizes the challenging task of attempting to determine the outcome of volcano unrest with only a few years or decades of volcano-monitoring data.

Similarly, in recent years, ground-based and space-based geodetic data (e.g., Interferometric Synthetic Aperture Radar (InSAR) and GPS), combined with seismic monitoring and other geophysical and geochemical investigations, have documented uplift and (or) subsidence at several other well-monitored long-dormant volcanic systems (e.g., Three Sisters volcanic center, Oregon; Long Valley Caldera, California; and Yellowstone Caldera, Wyoming-Montana). For more

information concerning these systems as well as other volcanic systems, the interested reader is referred to the following works: Wicks et al. (2002, 2006), Battaglia et al. (2003), Hill (2006), and Dzursin (2006 and chapters therein). It is important to stress that, while the level and intensity at these restless volcanic systems have varied within the periods of systematic instrumental monitoring (only a few decades at most), the available datasets and interpretations generally are inadequate to determine whether or not the ongoing unrest will culminate in eruption.

Kilauea Volcano, Hawaii (USA)

Brief mention should be made of frequently active volcanic systems – typically basaltic in composition – that are characterized by conspicuous and measurable unrest occurring nearly continuously over long time periods. Perhaps the best-studied volcanic systems exhibiting such behavior are those of Kilauea and Mauna Loa volcanoes on the Island of Hawaii (see Decker et al. 1987 and chapters therein; Tilling et al. 2010). These two volcanoes are among the most active in the world and have been well monitored since the establishment in 1912 of the Hawaiian Volcano Observatory, operated continuously by the USGS since 1948 (Babb et al. 2011). Abundant “baseline”

monitoring data, particularly for Kilauea, demonstrate well-defined patterns and outcomes of unrest (e.g., Fig. 5 in Tilling and Dvorak 1993; Fig. 8 in Tilling 2002). However, it must be emphasized that the behavior and outcomes of unrest observed for Kilauea – culminating either in subsurface intrusion only or in eruption – occur over much shorter time scales than those for the examples of explosive volcanoes discussed in this paper (e.g., Figs. 2, 3, and 8).

For Kilauea, because of its high eruption frequency and, at times, nearly continuous activity (e.g., Heliker et al. 2003), the outcomes of its volcano unrest and the need to issue early warnings are manageable for the civil authorities with little or no anxiety. This fortunate situation reflects a positive circumstance: because of the volcano's frequent and, at times, continuous eruptive activity, the dissemination of hazards information and early warnings is easily and routinely accomplished among the scientists, emergency-management officials, news media, and the affected public. Moreover, analyses of data for well-monitored, frequently active volcanoes, such as Kilauea, provide many case histories that can serve to test and refine volcano-monitoring techniques as well as eruption-forecasting methodologies. Nonetheless, it should be noted that eruptions at Kilauea and other basaltic volcanoes differ fundamentally from those of more explosive – hence, potentially more dangerous – strato-volcanoes that erupt more viscous materials of intermediate composition (e.g., Mount St. Helens, Mount Pinatubo).

Challenges in Achieving Refined Predictive Capability

The discussion in the preceding sections paints a somewhat discouraging picture: namely, possibly with very rare exceptions (e.g., the Mount St. Helens and Pinatubo case histories), to date scientists still lack a reliable capability for precisely *and* accurately predicting the outcomes of unrest at explosive volcanoes with low eruption frequency. This unsatisfactory state of affairs does not well serve societal needs to reduce the risks of

volcanic hazards and demands improvement. Considered below are some elements that I deem to be essential to achieve refined predictive capability and early warnings of possible eruptive activity.

More Geologic Mapping and Dating Studies of Volcanoes

Of the world's active and potentially active volcanoes, only a handful (mostly in the developed countries) have been studied sufficiently for the detailed and complete reconstruction of their past behavior, eruption frequencies, and associated hazards. Specifically, geologic mapping (at a scale of 1:25,000 or better) and dating studies of stratigraphically well-controlled eruptive products have been done for only a few volcanoes. Recorded historical eruptions for most volcanoes – especially those in the New World – rarely extend back in time for more than five centuries. Their eruptive histories generally are too short to serve as a comprehensive time-series dataset for quantitative statistical analysis of eruption frequency or repose intervals. Thus, it is essential to expand the time-series dataset by obtaining radiometric ages of prehistoric eruptive products. Moreover, in studies of eruption frequency, no matter how abundant the data used, the calculated frequencies obtained are necessarily *minima*, because the products of small eruptive events generally are lost to erosion and not preserved in the geologic record.

More Volcano Monitoring at More Volcanoes: Importance of Establishing Long-Term “Baseline” Behavior

As noted by Tilling (see p. 9 in Tilling 2008), most volcanologists believe that “...if a volcano is monitored extensively in real- or near-real time by well-deployed instrumental networks, it should be possible to make much better forecasts of the outcomes of volcano unrest.” Unfortunately, many of the world's active and potentially active volcanoes are poorly monitored, and some are not monitored at all. Thus, to advance our understanding of eruptive dynamics in order to make more precise early warnings, initiating or expanding volcano monitoring at many more volcanoes is

imperative. Ideally, volcano monitoring should be conducted in real- or near-real time, and it should involve a combination of monitoring techniques rather than reliance on any single one (Tilling 1995). Moreover, the longer the time span of volcano monitoring, the more diagnostic is the relevant database needed for detecting and interpreting any significant departure from the overall variation in the long-term “baseline” behavior of the volcano.

The Dilemma of “False Alarms”

With the current state of the art in volcanology and monitoring techniques, it is difficult, and generally impossible, for volcanologists to determine with any certainty: (1) the most likely outcome of onset of unrest at a long-dormant volcano and (2) the eventual possible occurrence of a major eruption in the course of long-duration but relatively low-level ongoing unrest. Nonetheless, scientists still have an obligation to emergency managers and the threatened populations to provide the best-possible hazards information, forecasts or predictions, and warnings within the limitations of available knowledge, technology, and volcano-monitoring techniques. In so doing, the scientists run the unavoidable risk of raising a “false alarm” and the associated criticism and loss of credibility when the forecasted event fails to materialize. On the other hand, emergency-management officials as well as the general public must be prepared to accept the disruptive socioeconomic consequences and costs of occasional “false alarms.” The Rabaul case history (previously discussed in section “[Rabaul Caldera \(Papua New Guinea\)](#)”) underscores the sentiment of Banks et al. (see p. 78 in Banks et al. 1989): “...False alarms themselves can provide, through objective assessment of the scientific and public response to a volcanic crisis that ended without eruption, valuable lessons useful ... for the next crisis, which could culminate in an eruption.” Even so, every attempt should be made to minimize “false alarms” to the extent possible, in order to maintain scientific credibility and the confidence of the emergency-management authorities and the populace at risk.

Future Directions

An obvious critical need is the comprehensive geoscience studies of more volcanoes, to reconstruct their prehistoric eruptive histories, thereby extending the time-series database needed for quantitative analysis and interpretation. In addition, volcano monitoring, even if rudimentary – deploying only 1–3 seismometers and making simple field observations and measurements (Swanson 1992) – is needed for many more volcanoes, both to establish “baselines” and to provide early warning. Obviously, future research necessarily should also be focused on methodologies to make more robust the acquisition and processing of real- or near-real-time volcano-monitoring data, using or refining *existing* instrumentation and techniques, volcano databases, analytical tools, and empirical and theoretical models. An especially promising avenue of future research would require the wider utilization of broadband instruments in seismic-monitoring networks at active and potentially volcanoes. This would enable systematic and quantitative analysis and interpretation of long-period (LP) and very-long-period (VLP) seismicity, which from worldwide experience precedes and accompany nearly all eruptions ► “[Volcanoes, Non-linear Processes in.](#)”

A long-standing and vexing problem for volcanologists is that, at present, no geophysical or geochemical criteria exist to identify the distinguishing characteristics between volcano unrest that merely ends in a subsurface intrusion of magma and unrest that culminates with magma breaking the surface to produce an eruption. From the vantage point of the emergency-management officials, subsurface intrusion or movement of magma, while of great academic interest to volcanologists in deciphering volcano dynamics, poses no threat to public safety. In contrast, the threat of an imminent eruption requires timely decisions and mitigative actions to ensure public safety. Thus, the future development, if possible, of quantitative criteria to distinguish the precursory pattern of a subsurface intrusion from that for an eruption would mark a quantum leap in the young science of volcanology. In this regard, an

improved understanding of magma intrusion versus eruptive dynamics might provide scientists more powerful tools in choosing between the various paths shown in Fig. 3 as the most likely outcome of volcano unrest.

In attempting to quantify estimates of the probabilities of specified hazardous volcanic events within specified time frames (i.e., eruption predictions), scientists have made increasing use of “materials failure” models (e.g., Murray and Ramirez Ruiz 2002; Voight 1988; Voight and Cornelius 1991), “event trees” (e.g., Newhall and Hoblitt 2002), and more statistically rigorous variants of the “event tree,” “decision tree,” “pattern recognition,” and “occurrence pattern” approaches (e.g., Marzocchi 1996; Marzocchi et al. 2004, 2008; Mulargia et al. 1991; Sandri et al. 2004) ► **“Volcanic Eruptions: Stochastic Models of Occurrence Patterns”**). Depending on the particular circumstances of the volcano (es) and eruptions under study, these and other statistical tools have their individual strengths and weaknesses. Taken together, however, the utility and potential successful application of such approaches are inherently hampered by the incompleteness and (or) too short time span of the eruptive history or volcano-monitoring datasets analyzed. Thus, to become more robust, the probabilistic methodologies must employ larger and more complete datasets. In any case, while useful, the existing techniques or models cannot yield results with the precision and accuracy that emergency managers demand and need as they confront a “decision window” during a volcanic crisis. Simply stated, even if the calculated results prove to be accurate, their error bars generally exceed the time span bracketed by the “decision window.”

In recent years, the World Organization of Volcano Observatories (WOVO) – a Commission of the International Association of Volcanology and Chemistry of the Earth’s Interior – has launched an ambitious but to date poorly funded effort, called *WOVOdat*, to construct a global database of empirical data on volcanic unrest. This database is being developed as a user-friendly Internet-based system with query and analysis tools (Schwandner and Newhall 2005). Even

though a basic design and schema now exist for the construction of *WOVOdat* (Venezky and Newhall 2007), the actual compilation and integration of data are just barely beginning. Since the late 2000s, the *WOVOdat* project has received some financial support from the Singapore Government and Nanyang Technological University (Newhall, 11 February 2008, “written communication”). This favorable development brightens the prospects for it ultimately to become a powerful tool in evaluating patterns and possible outcomes of volcanic unrest. However, as of this writing (June 2014), adequate stable funding for *WOVOdat* is still not yet secured (Newhall, 14 June 2014, “written communication”). Thus, until *WOVOdat* becomes fully operational, we must continue to rely on the available datasets – however imperfect and region specific – and to employ and refine the existing analytical methodologies and statistical models.

Clearly, continuing scientific studies are needed to advance volcanology and to better characterize and understand eruptive phenomena and associated hazards. It should be remembered, however, that scientific data – no matter how opportune or precise – must also be communicated effectively to, and acted upon in a timely manner by, emergency-management authorities (Tilling 1995). In so doing, some attributes of an effective volcano warning are crucial. As emphasized by Newhall (see p. 1186 in Newhall 2000), for a warning to succeed, it must be: “. . . accurate, . . . understandable, . . . timely, . . . widely disseminated, . . . credible, . . . and catalytic.” By “catalytic,” Newhall means that the warning should contain “clear suggestions” about what actions the recipients can take to avoid or mitigate the threat, as well as “convincing reasons” why these actions are preferable to inaction. Lessons learned from some notable volcanic disasters in recent decades (e.g., El Chichón, Mexico, 1982 (Tilling 2009); Nevado del Ruiz, Colombia, 1985 (Voight 1990)) have prompted the global volcanologic community to devise new volcano warning systems, or improve existing systems, to try to make them more effective (e.g., Fearnley 2013; Garcia et al. 2014; Garcia and Fearnley 2012). There is now an awareness that improvements in

warnings can only be achieved with close collaboration between volcanologists and specialists in the social sciences and other disciplines (e.g., Donovan and Oppenheimer 2014; Donovan et al. 2012; Fearnley 2013; Garcia et al. 2014; Garcia and Fearnley 2012; Haynes et al. 2008). Volcanic hazards associated with powerful eruptions are not necessarily confined within national political boundaries and can have global impact. As mentioned earlier, drifting volcanic ash clouds during the April 2010 Eyjafjallajökull eruption crisis severely disrupted international civil aviation. The human aggravation and economic loss caused by this crisis could have been substantially reduced if international coordination and regulation of European air space had existed at the time (Alexander 2013 and references therein). It is instructive and sobering to remember that the Eyjafjallajökull eruption was moderate in size and of relatively short duration, compared to some other large historical eruptions. Ideally, a truly effective early warning system should be regional – or preferably global – in design, scope, and mandate.

Acknowledgments An earlier draft of this article benefited greatly from constructive reviews and helpful suggestions by L. J. Patrick Muffler and Fred Klein (both of the US Geological Survey, Menlo Park). To them, I offer them my sincere thanks. The views expressed in this chapter have been shaped by my personal involvement in responses to several of the volcanic crises in recent decades and by enlightening and instructive interactions and discussions with many colleagues in the global volcanological community.

Bibliography

Primary Literature

- Alexander A (2013) Volcanic ash in the atmosphere and risks for civil aviation. *Int J Disaster Risk Sci* 4:9–19. doi:10.1007/s13753-013-0003-0
- Babb JL, Kauahikaua JP, Tilling RI (2011) The story of the Hawaiian Volcano Observatory – a remarkable first 100 years of tracking eruptions and earthquakes: U.S. Geological Survey General Information Product 135, 60 p. Available at <http://pubs.usgs.gov/gip/135/>
- Banks NG, Tilling RI, Harlow DH, Ewert JW (1989) Volcano monitoring and short-term forecasts. In: Tilling RI (ed) *Short courses in geology*, vol 1, Volcanic hazards. American Geophysical Union, Washington, DC, pp 51–80 (Chap 4)
- Battaglia M, Roberts C, Segall P (2003) The mechanics of unrest at Long Valley caldera, California: 2. Constraining the nature of the source using geodetic and micro-gravity data. *J Volcanol Geotherm Res* 127:219–245
- Berrino G, Corrado G, Luongo G, Toro B (1984) Ground deformation and gravity changes accompanying the 1982 Pozzuoli uplift. *Bull Volcanol* 47:187–200
- Blong R, McKee C (1995) The Rabaul eruption 1994: destruction of a town. Natural Hazards Research Centre, Macquarie University, Sydney, 52 pp
- Casadevall TJ (ed) (1994) Volcanic ash and aviation safety: proceedings of the first international symposium on volcanic ash and aviation safety. US Geological Survey Bulletin 2047. Government Printing Office, Washington, DC, 450 pp
- Chouet B (2004) Volcano seismology. *Pure Appl Geophys* 160:739–788
- Crandell DR, Mullineaux DR, Rubin M (1975) Mount St. Helens volcano: recent and future behaviour. *Science* 187:438–441
- Decker RW, Wright TL, Stauffer PH (eds) (1987) Volcanism in Hawaii, vol 1 and 2, US Geological Survey professional paper 1350. U.S. Government Printing Office, Washington, DC, 1667 pp
- De Vito M, Lirer L, Mastrolorenzo G, Rolandi G (1987) The 1538 Monte Nuovo eruption (Campi Flegrei, Italy). *Bull Volcanol* 49:608–615
- Donovan A, Oppenheimer C (2014) Science, policy and place in volcanic disasters: insights from Montserrat. *Environ Sci Policy* 39:150–161
- Donovan A, Oppenheimer C, Bravo M (2012) Social studies of volcanology: knowledge generation and expert advice on active volcanoes. *Bull Volcanol* 74:677–689
- Dzurisin D (ed) (2006) Volcano deformation: geodetic monitoring techniques. Springer-Praxis, Berlin, 441 pp
- Fearnley CJ (2013) Assigning a volcano alert level: negotiating uncertainty, risk, and complexity in decision-making processes. *Environ Plann* 45:1891–1911. doi:10.1068/a4542
- Garcia C, Fearnley C (2012) Evaluating critical links in early warning systems for natural hazards. *Environ Hazard* 11:123–137
- García A, Berrocoso M, Marrero JM, Fernández-Ros A, Prates G, De la Cruz-Reyna S, Ortiz R (2014) Volcanic alert system (VAS) developed during the 2011–2014 El Hierro (Canary Islands) volcanic process. *Bull Volcanol* 76. doi:10.1007/s00445-014-0825-7
- Guffanti M, Thomas J, Casadevall TJ, Budding K (2010) Encounters of aircraft with volcanic ash clouds: a compilation of known incidents, 1953–2009. U.S. Geological Survey Data series 545, ver 1.0, 12 pp, plus 4 appendixes including the compilation database. Available at <http://pubs.usgs.gov/ds/545>
- Haynes K, Barclay J, Pidgeon N (2008) Whose reality counts? Factors affecting the perception of volcanic risk. *J Volcanol Geotherm Res* 172:259–272
- Heliker C, Swanson DA, Takahashi TJ (eds) (2003) The Pu'u O'o-Kupaianaha eruption of Kilauea Volcano,

- Hawaii: the first 20 years, US Geological Survey professional paper 1676. U.S. Geological Survey, Reston, 206 pp
- Hill DP (2006) Unrest in Long Valley Caldera, California, 1978–2004. *Geol Soc Lond* 269:1–24
- ISDR (2004) Living with risk: international strategy for disaster reduction (ISDR), vol 1, 431 pp and vol 2, 126 pp. United Nations, New York/Geneva
- Klein F (1982) Patterns of historical eruptions at Hawaiian volcanoes. *J Volcanol Geotherm Res* 12:1–35
- Klein F (1984) Eruption forecasting at Kilauea Volcano, Hawaii. *J Geophys Res* 89(B5):3059–3073
- Lipman PW, Mullineaux DR (eds) (1981) The 1980 eruptions of Mount St. Helens, Washington. US Geological Survey Professional Paper 1250. U.S. Government Printing Office, Washington, DC, 844 pp
- Marzocchi W (1996) Chaos and stochasticity in volcanic eruptions the case of Mount Etna and Vesuvius. *J Volcanol Geotherm Res* 70:205–212
- Marzocchi W, Sandri L, Gasparini P, Newhall C, Boschi E (2004) Quantifying probabilities of volcanic events: the example of volcanic hazards at Mt. Vesuvius. *J Geophys Res* 109, B11201. doi:10.1029/2004JB003155
- Marzocchi W, Sandri L, Selva J (2008) BET_EF: a probabilistic tool for long- and short-term eruption forecasting. *Bull Volcanol* 70(5):623–632. <http://dx.doi.org/>
- McKee CO, Johnson RW, Lowenstein PL, Riley SJ, Blong RJ, de St. Ours P, Talai B (1985) Rabaul caldera, Papua New Guinea: volcanic hazards, surveillance, and eruption contingency planning. *J Volcanol Geotherm Res* 23:195–237
- McNutt SR (1996) Seismic monitoring and eruption forecasting of volcanoes: a review of the state of the art and case histories. In: Scarpa R, Tilling RI (eds) *Monitoring and mitigation of volcano hazards*. Springer, Heidelberg, pp 99–146
- McNutt SR (2000a) Seismic monitoring. In: Sigurdsson H, Houghton B, McNutt SR, Rymer H, Stix J (eds) *Encyclopedia of volcanoes*. Academic, San Diego, pp 1095–1119 (Chap 68)
- McNutt SR (2000b) Synthesis of volcano monitoring. In: Sigurdsson H, Houghton B, McNutt SR, Rymer H, Stix J (eds) *Encyclopedia of volcanoes*. Academic, San Diego, pp 1167–1185 (Chap 71)
- Mulargia F, Gasparini P, Marzocchi W (1991) Pattern recognition applied to volcanic activity: identification of the precursory patterns to Etna recent flank eruptions and periods of rest. *J Volcanol Geotherm Res* 45:187–196
- Murray JB, Ramirez Ruiz JJ (2002) Long-term predictions of the time of eruptions using remote distance measurement at Volcán de Colima, México. *J Volcanol Geotherm Res* 117(1–2):79–89
- Newhall CG (2000) Volcano warnings. In: Sigurdsson H, Houghton B, McNutt SR, Rymer H, Stix J (eds) *Encyclopedia of volcanoes*. Academic, San Diego, pp 1185–1197 (Chap 72)
- Newhall CG, Punongbayan RS (eds) (1996) Fire and mud: eruptions and lahars of the Mount Pinatubo, Philippines. Philippine Institute of Volcanology and Seismology, Quezon City/University of Washington Press, Seattle, 1126 pp
- Newhall CG, Hoblitt RP (2002) Constructing event trees for volcanic crises. *Bull Volcanol* 64:3–20
- OFCM (2004) Proceedings of 2nd international conference on volcanic ash and aviation safety, 21–24 June 2004. U.S. Department of Commerce/National Oceanic and Atmospheric Administration, Alexandria. <http://www.ofcm.gov/ICVAAS/Proceedings2004/ICVAAS2004-Proceedings.htm>. Accessed 15 June 2014
- Osservatorio Vesuviano (2014) Campi Fleigrei geodesy. <http://www.ov.ingv.it/ov/it/campi-flegrei/attivita-recente.html>. Accessed 14 June 2014
- Sandri L, Marzocchi W, Zaccarelli L (2004) A new perspective in identifying the precursory patterns of eruptions. *Bull Volcanol* 66:263–275
- Schwandner FM, Newhall CG (2005) WOVodat: the World Organization of Volcano Observatories database of volcanic unrest. *Eur Geosci Union, Geophys Res Abstr* vol 7, abstract # 05-J-09267 (extended abstract), 2 pp
- Sherrod DR, Scott WE, Stauffer, PH (eds) (2008) A volcano Rekindled: the renewed eruption of Mount St. Helens, 2004–2006. U.S. Geological Survey, Reston, Virginia, 856 pp
- Siebert L, Simkin T, Kimberly P (2010) *Volcanoes of the world*, 3rd edn. Smithsonian Institution, Washington DC/University of California Press, Berkeley, 568 pp
- Sigurdsson H, Houghton B, McNutt SR, Rymer H, Stix J (eds) (2000) *Encyclopedia of volcanoes* (and chapters therein). Academic, San Diego, 1417 pp
- Sparks RSJ (2003) Frontiers: forecasting volcanic eruptions. *Earth Planet Sci Lett* 210(1–2):1–15
- Swanson DA (1992) The importance of field observations for monitoring volcanoes, and the approach of “keeping monitoring as simple as practical”. In: Ewert JW, Swanson DA (eds) *Monitoring volcanoes: techniques and strategies used by the staff of the Cascades Volcano Observatory, 1980–90*, vol 1966, US Geological Survey bulletin. U.S. Government Printing Office, Washington, DC, pp 219–223
- Swanson DA, Casadevall TJ, Dzurisin D, Holcomb RT, Newhall CG, Malone SD, Weaver CS (1985) Forecasts and predictions of eruptive activity at Mount St. Helens, USA: 1975–1984. *J Geodyn* 3:397–423
- Tilling RI (1995) The role of monitoring in forecasting volcanic events. In: McGuire WJ, Kilburn CRJ, Murray JB (eds) *Monitoring active volcanoes: strategies, procedures, and techniques*. UCL Press, London, pp 369–402
- Tilling RI (2000) Mount St. Helens 20 years later: what have we learned? *Geotimes* 45:14–19
- Tilling RI (2002) Volcanic hazards. In: Meyer RA (ed) *Encyclopedia of physical science and technology*, vol 17, 3rd edn. Academic, San Diego, pp 559–577
- Tilling RI (2003) Volcano monitoring and eruption warnings. In: Zschau J, Küppers AN (eds) *Early warning systems for natural disaster reduction*. Springer, Berlin, pp 505–510 (Chap 5)
- Tilling RI (2005) Volcano hazards. In: Marti J, Ernst G (eds) *Volcanoes and the environment*. Cambridge University Press, Cambridge, pp 56–89 (Chap 2)

- Tilling RI (2008) The critical role of volcano monitoring in risk reduction. *Adv Geosci* 14:3–11
- Tilling RI (2009) El Chichón's "surprise" eruption in 1982: lessons for reducing volcano. *Geofis Int* 48:3–19
- Tilling RI, Dvorak JJ (1993) The anatomy of a basaltic volcano. *Nature* 363:125–133
- Tilling RI, Topinka L, Swanson DA (1990) Eruptions of Mount St. Helens: past, present, and future, 2nd edn. U.S. Geological Survey, Reston, Virginia, 56 pp
- Tilling RI, Heliker C, Swanson DA (2010) Eruptions of Hawaiian volcanoes: past, present, and future. U.S. Geological Survey, Reston, Virginia, 63 pp
- Troise C, De Natale G, Pingue F, Obrizzo F, De Martino P, Tammaro U, Boschi E (2007) Renewed ground uplift at Campi Flegrei caldera (Italy): new insight on magmatic processes and forecast. *Geophys Res Lett* 34, L03301. doi:10.1029/2006GL28545
- Turner MB, Cronin SJ, Bebbington MS, Platz T (2008) Developing probabilistic eruption forecasts for dormant volcanoes: a case study from Mt. Taranaki, New Zealand. *Bull Volcanol* 70:507–515
- Venezky DY, Newhall CG (2007) WOVODat design document: the schema, table descriptions, and create table statements for the database of worldwide volcanic unrest (WOVODat version 1.0). US Geological Survey Open-File report 2007-1117, 184 pp
- Voight B (1988) A method for prediction of volcanic eruptions. *Nature* 332:125–130
- Voight B (1990) The 1985 Nevado del Ruiz Volcano catastrophe: anatomy and retrospection. *J Volcanol Geotherm Res* 42:151–188
- Voight B, Cornelius RR (1991) Prospects for eruption prediction in near real-time. *Nature* 350:695–698
- WG99, Working Group on California Earthquake Probabilities (1999) Earthquake probabilities in the San Francisco Bay region: 2000 to 2030 – a summary of findings. US Geological Survey Open-File report 99-517, 60 pp
- Wicks C, Dzurisin D, Ingebritsen S, Thatcher W, Lu Z, Iverson RM (2002) Magmatic activity beneath the quiescent Three Sisters volcanic center, central Oregon Cascade Range, USA. *Geophys Res Lett* 29(7): 26–751, 26–754. doi:10.1029/2001GL014205
- Wicks C, Thatcher W, Dzurisin D, Svarc J (2006) Uplift, thermal unrest, and magmatic intrusion at Yellowstone Caldera. *Nature* 440:72–75

Books and Reviews

- Chester D (ed) (1993) *Volcanoes and society*. Edward Arnold (a Division of Hodder & Stoughton), London, 351 pp
- Heiken G (2013) From Kilauea Iki 1959 to Eyjafjallajökull 2010: how volcanology has changed! *Geol Soc Am Spec Pap* 500:33–63
- Lockwood JP, Hazlett RW (2010) *Volcanoes: global perspectives*. Wiley-Blackwell, Chichester, 541 pp
- Zoback ML, Geist E, Pallister J, Hill DP, Young S, McCausland W (2013) *Advances in natural hazard science and assessment, 1963–2013*. *Geol Soc Am Spec Pap* 501:81–154
- Marti J, Ernst G (eds) (2005) *Volcanoes and the environment*. Cambridge University Press, Cambridge, 471 pp
- Scarpa R, Tilling RI (eds) (1996) *Monitoring and mitigation of volcano hazards*. Springer, Heidelberg, 841 pp
- Schmincke H-U (2005) *Volcanism*. Springer, New York, p 324

Volcanic Hazards Warnings: Effective Communications of

C. J. Fearnley

Department of Science and Technology Studies,
University College London, London, UK

Article Outline

Glossary

Introduction: Challenges to Volcanic Crisis

Communication

Early Warning Systems

Volcano Warnings

The Emergence and Challenge of Standardization

Future Directions

Bibliography

Glossary

CAS Complex Adaptive System

EWS Early Warning System

IDNDR International Decade for Natural Disaster Reduction

ISDR International Strategy for Disaster Reduction

NIMS National Incident Management System

NOTAM Notice to Airmen

NVEWS National Volcano Early Warning System

USGS US Geological Survey

VAA Volcanic Ash Advisory

VAAC Volcanic Ash Advisory Center

VALS Volcano Alert Level System

VAN Volcanic Activity Notice

VEWS Volcano Early Warning System

VHP Volcano Hazard Program

VONA Volcano Observatory Notices for Aviation

Introduction: Challenges to Volcanic Crisis Communication

There have been numerous volcanic crises and disasters over the last 100 years, despite

significant advances in scientific knowledge, monitoring technologies, and the development of statistical and GIS-based models. Often these disasters have resulted from a breakdown in communication that has often been the result of no, or poor, early warning systems in place, or the failure of society to respond effectively. One such tragic example is the 1985 eruption of Nevado del Ruiz in Colombia which, despite having only a volcanic explosivity index (VEI) of 3, killed over 23,000 people in the city of Armero, when a fatal lahar that was forecast and expected by scientists traveled over 70 km to strike the city (Hall 1990). There was no formal early warning system in place; however, hazard maps had been drawn by the Colombian Geological and Mine Institute (INGEOMINAS) and United States Geological Survey (USGS), with scientists identifying numerous prior lahars that had traveled through Armero. The resulting high death toll was because local authorities and communities did not act on warnings, as Voight eloquently states “the catastrophe was not caused by technological ineffectiveness or defectiveness, nor by an overwhelming eruption, or by an improbable run of bad luck, but rather by cumulative human error – by misjudgment, indecision and bureaucratic shortsightedness. Armero could have produced no victims, and therein dwells its immense tragedy” (1990, p. 349). This tragedy, among many others, highlights the important role of effective communications in reducing risks posed by volcanic hazards.

The Diversity of Volcanic Hazards

Volcanic hazards are complex because of the diversity in the types of hazards that a volcano can produce, both primary and secondary, and the differing geographical locations that these hazards impact, and the differing time scales (for an extensive overview see Sigurdsson et al. 2015). For example, volcanic ash generally only occurs during an explosive eruption, and can affect a large area downwind of the volcano as the ash travels through the atmosphere and is deposited. In

contrast, lava flows only impact areas within relatively close proximity of an erupting volcano, but these flows can occur over decades during a prolonged eruption. A nonerupting volcano can still generate lahars (particularly if there is a storm, heavy rainfall, or glacial melt), landslides (due to unstable slopes), and subsequently tsunamis (displacement of water in or near to the volcano). Volcanoes produce the most diverse range of hazards of any phenomenon, which can lead to extensive secondary hazards such as crop failure, famine, disease, contamination, and climate change that can result in more fatalities and socioeconomic impact than the primary hazards of a volcano (Oppenheimer 2011). It is this diversity and complexity of the hazard that creates problems when trying to communicate potential hazards and risks. In addition, volcanoes are complex phenomenon, a nondeterministic emergent system, making them exceedingly challenging to predict, but there is however scope to forecast potential activity (Kilburn 2003; Sparks and Aspinall 2004). Forecasting and prediction are often considered synonymous; however, it is recommended to adopt the following definitions. A forecast is a comparatively imprecise statement of the time, place, and nature of expected activity. Prediction is a comparatively precise statement of the time, place, and ideally, the nature and size of impending activity (Swanson et al. 1985). With complex phenomenon that is not easily forecast, a wide range of generated hazards, there is also the additional risk that society generates by building and living near or even distal to volcanoes.

The communication and negotiation of potential hazards and their impact on society is typically classified as Disaster Risk Reduction measures (Kelman and Kelman 2017; Wisner et al. 2012). There are numerous preparedness and mitigation strategies that can be put into place. Mitigation strategies are policies or procedures that lead to preplanned actions that operate before or during a hazard event to reduce its impact on vulnerable populations. Common examples include land-use and development planning; engineering strategies such as tsunami barriers, river or tidal flood defenses, and seismically resilient buildings; and warning systems that foster education, evacuation plans, and communication to enable mitigation

actions at the time of hazard events or in anticipation of them. Different mitigation strategies require differing timescales and methods of implementation according to both the nature of the hazards and the vulnerabilities of the exposed societies. Mitigation systems consist of all the mitigation strategies implemented together to ensure “the lessening or limitation of the adverse impacts of hazards and related disasters” (UNISDR 2007). Day and Fearnley (2015) divide mitigation strategies into three classes according to the timing of the actions that they prescribe. Permanent mitigation strategies prescribe actions such as construction of tsunami barriers or land-use restrictions: they are frequently both costly and “brittle” in that the actions work up to a design limit of hazard intensity or magnitude and then fail. Responsive mitigation strategies prescribe actions after a hazard source event has occurred, such as evacuations, that rely on capacities to detect and quantify hazard events and to transmit warnings fast enough to enable at risk populations to decide and act effectively. Anticipatory mitigation strategies prescribe use of the interpretation of precursors to hazard source events as a basis for precautionary actions, but challenges arise from uncertainties in hazard behavior. Day and Fearnley’s classification provides an insight into the challenges of the temporal dynamics of natural hazards that constrain the ability to provide warnings and key parameters surrounding potential events and hazards. It is these timing constraints that shape the actions prescribed by a strategy, and the adaptability of these actions to individual hazard events, both of which shape the types of warnings and communication that are possible for volcanic hazards.

Volcano Mitigation and Communication Strategies

For volcanoes, most mitigation strategies are anticipatory. Even with permanent structures in place such as SABO dams (commonly found in Japan), should an eruption be imminent, populations are still likely to be evacuated. Anticipatory mitigation strategies prescribe use of the interpretation of precursors to hazard source events as a basis for precautionary actions, but challenges arise from uncertainties in hazard

behavior. Societies also struggle to manage precautionary approaches as the cost of evacuations is high, many people do not want to leave their home or their businesses or farms as the cost is great, particularly for extended times (Tobin and Whiteford 2002; Stirling 2007). Therefore, either greater certainty is needed or greater support from the government to rehouse and support populations for the time as required. To try and negotiate these differing needs given the inherent uncertainties, Early Warning Systems (EWS) are usually implemented as a cost-beneficial tool to manage the precautionary and anticipatory issues that emerge from volcanic crises.

EWS are a key component of mitigation strategies, whereby communication and decision-making based upon information provided by monitoring and warning technologies leads to actions in response to that information. Despite an abundance of EWS-related research, there is little consensus about what they are or how they are defined (Glantz 2004). EWS are seen as “a means of getting information about an impending emergency, communicating that information to those that need it, and facilitating good decisions and timely response by people in danger” (Mileti and Sorenson 1990, pp. 2–1). While this is a simple definition, the operation of an EWS is far more complex, partly due to variations spatially (global, national, regional, local), temporally (rapid onset, slow onset, frequent, infrequent), in function (safety, property, environment), and in hazard (weather, climate, geo-hazard). EWS also operate in different economic, political, and social circumstances; use different communicative tools (from technology to word of mouth); and link many different organizations (or actors) such as science (government and private), engineering, technology, government, news/media, and the public. This leads to different perspectives of what EWS are, and what they should do. For governments, EWS are an important tool for disaster-risk reduction (DRR) measures; consequently, EWS tend to be highly centralized. Decisions have to be made about the benefits of EWS relating to: cost-benefit, timeliness (what constitutes a warning; are they a forecast, projection, or trend; and how early is early), establishing different levels of warning, and lastly accountability.

As part of the warning process, there are a number of key tools used by volcano observatories and institutions globally that communicate during times of escalated unrest. These include various written statements, alert level systems, notification systems, and the use of social media. These are explored further in this chapter, but it is important to note that there are a diverse and wide range of communication tools that can be used during volcanic crises. These range from (and can all be found in Fearnley et al. 2017b): understanding historical texts and knowledge to help prepare for future events; developing community knowledge and DRR practices that embrace oral and indigenous knowledges; stakeholder engagement and communication that recognizes cultural, disciplinary, and economic divides; decision-making tools; the use of statistic (particularly Bayesian) models to forecast volcanic unrest and activity; the role of insurance to mitigate against losses; the use of hazard and risk maps; the use of satellite imagery and geo-spatial technologies inform real-time analysis, communication, and decision-making; the role of art and performance in communicating volcanic hazards and risks, alongside folklore; the role of education for all ages in various teaching environments; and the role of social media. While these approaches are all vital to modern day volcanic crisis communication, only some aspects of these approaches contribute to the development of volcano hazard warnings. This chapter subsequently aims to examine the role of early warning systems, various types of volcano hazard warnings, the challenges of standardization focusing on a case study of the United States Geological Survey’s Volcano Alert Level Systems that were standardized in 2006, and the value of complexity approaches when considering what is effective communication for volcano hazard warnings.

Early Warning Systems

Early warning systems (EWS) have existed for a number of hazards, but the most significant and oldest system is the Pacific Tsunami Warning System, established in 1949 following the 1946 Aleutian Island earthquake that generated a

tsunami that killed 165 people in both Alaska and Hawaii. Despite growing populations near volcanoes, little research has been devoted to establishing best practice for volcano EWS to minimize loss of life and socioeconomic damage prior to and during volcanic crises.

Generating effective warnings is particularly challenging for several key reasons. First, volcanologists and related scientists are still developing theories to understand the origin, processes, and eruptive behavior of volcanoes and their numerous associated hazards. Second, volcanic hazards occur within different social contexts involving different cultures, and economic and political circumstances. In addition, volcanic activity tends to occur over long time frames relative to human time-scales and, in particular, periods of political office, and therefore are not normally a political priority. This commonly results in limited funding and resources for research and volcano observatory upkeep, leading to limited volcanic hazard awareness. Finally, institutional influences can lead to increasing levels of bureaucracy so that decisions become complex and take a long time to make. Managing a volcanic crisis can involve numerous institutions from emergency managers/civil protection, to weather services, to land managers and owners, to the media, to businesses and key infrastructure, to the public, making it difficult to maintain effective communications, both internally and externally.

To devise effective EWS, there is a need to consider what these systems are and how they have evolved over time, what they do, how they operate, and how their success can be measured. These aspects are addressed below.

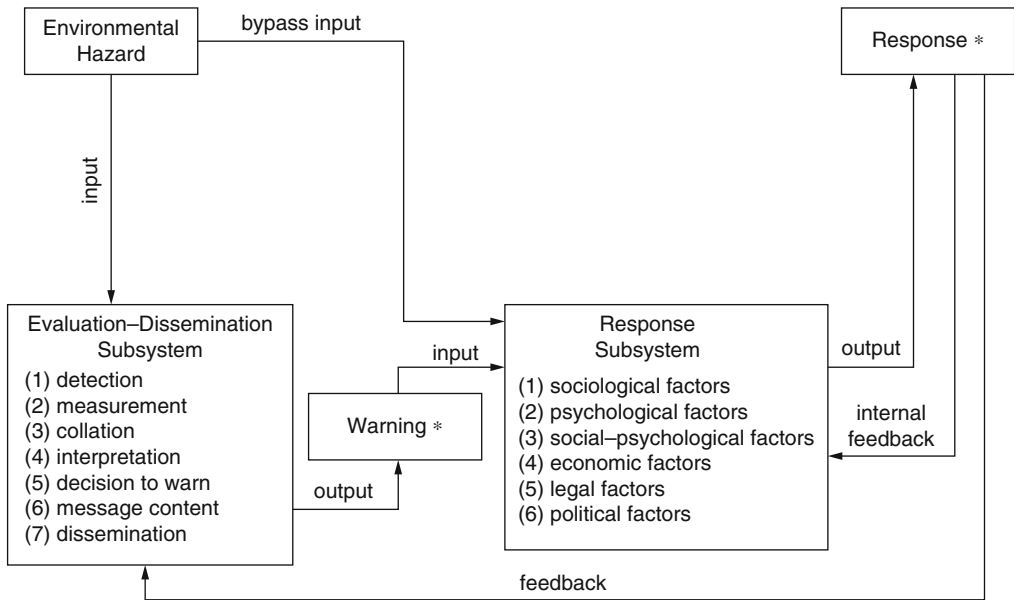
An Evolution from the Linear to the Complex

The diversity in EWS and of the agencies involved results in different ways of conceptualizing EWS. This section explores theoretical approaches to EWS by reviewing the concept and its evolution from a linear to a complex system.

EWS form a relatively new area of inquiry within the context of disaster research, obtaining growing recognition in 1960s. Prior to this, studies typically viewed early warning as a linear process (Gillespie and Perry 1976), where there

is a clear relationship between a hazard occurring and generating a warning, forming a cause and effect relationship. Linear processes are characteristically embodied in the Newtonian paradigm, often referred to as reductionist, but in 1969 a study by Barton changed this to view early warnings as a “system.” Barton’s work (1969) on disaster classification generated a paradigm shift from the descriptive to the analytical, by developing four classifying variables in his typology of disasters: scope of impact, speed of onset, duration of impact, and social preparedness. This work influenced Gillespie and Perry who said that “by adopting a systems perspective, the disaster researcher can not only describe and classify disasters more effectively, but can also move towards a more analytic approach” (Gillespie and Perry 1976, p. 305). A systemic approach enabled the development of models for the prediction of individual, group, and organizational behaviors, going beyond the simplistic cause and effect relationships within an early warning.

Although the idea of “systems” influence on disasters had first been identified in 1958 by Form and Nostow, it took decades to take hold. General Systems Theory (GST) emerged following the Second World War as an interdisciplinary approach to the field of science and the study of the complex systems in nature and society (Bertalanffy 1975). The term “systems” has many definitions, although the one adopted here is of a group of interacting, interrelated, or interdependent elements forming a complex whole, which is nearly always defined with respect to a specific purpose (Kim 1994). Originating in biological studies in the 1920s, GST recognized that systems are greater than the sum of their parts (Bertalanffy and Woodger 1933), providing a holistic approach to disaster studies by demonstrating that the processes involved are interrelated. In 1975, models of idealized EWS were developed such as in Fig. 1, which rather than showing an EWS as a linear progression through the different stages of disasters in chronological order, indicated that an EWS comprised of subsystems (in this case evaluation-dissemination and response) that have inputs, outputs, and feedback between them.



* Major dependent variables in the system.

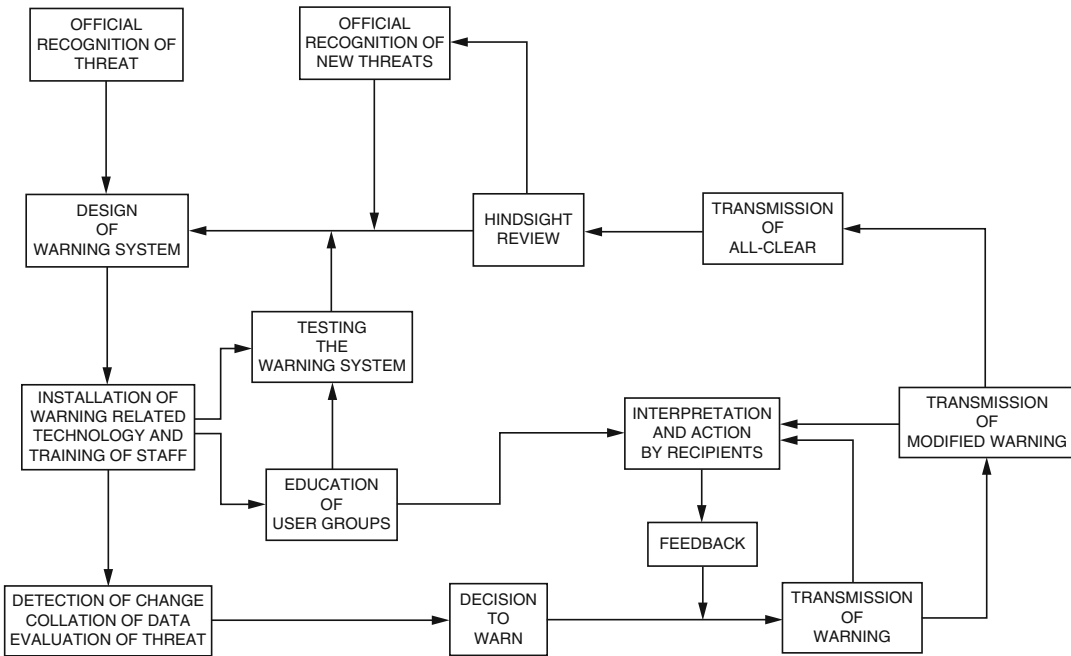
Volcanic Hazards Warnings: Effective Communications of, Fig. 1 Systems model of a warning system (White and Haas 1975, p. 185)

By the 1980s, Foster (1980) identified that decision-making and communication processes between different actors in EWS were nonlinear and could be understood better within the context of systems theory as a dynamic system. Foster also developed an idealized EWS to represent the different stages, using a system style layout as seen in Fig. 2. Although this model recognized the role of organizations and policy, it maintains an element of linearity rather than presenting a series of feedback loops that are multidirectional enabling a systems approach, as Foster states “every warning system should be designed to facilitate a *two-way flow* of information” (Foster 1980, p. 203) (author’s emphasis).

The EWS models developed by White and Haas (1975) and Foster (1980) (in Figs. 1 and 2 respectively) divide EWS into component parts and consider each part separately to ensure their proper function (White 1995). While these models are idealized and are not descriptive of what actually happens in an EWS, the models also struggle to view EWS as a system because they fail to “identify emergent properties arising from interacting elements and because it does not

consider that the behaviour of systems is due as much to their external environment as to their internal mechanisms” (White 1995, p. 41). White argues that disaster studies tools that provide a holistic approach, by considering how human behavior and context can affect the management of risk, should be used.

By the late 1990s, there was growing recognition that interactions between natural environments, human perception, actions, and organizations are part of a genuinely “complex” system (Mileti 1999). The term “complex” has become a popular and often misused term both in the physical and social sciences. The complex systemic approach “focuses on interaction among the elements of a system and on the effects of its interactions; it examines a variety of factors at one time; it integrates time, feedback, and uncertainty” (Mileti 1999, p. 107). It is the reciprocal interactions or feedback among variables or subsystems, as well as time delays in seeing the results, that create complexity, making the system difficult to understand (Senge 1990). As a result, complexity highlights serious limitations to our scientific knowledge because it breaks traditional



Volcanic Hazards Warnings: Effective Communications of, Fig. 2 Idealized warning systems (Foster 1980, p. 172)

reductionist Newtonian thinking that regards science as infinitely divisible and measurable (Capra 1996); “complexity argues against reductionism, against reducing the whole to the parts” (Urry 2005a, p. 401). The concept of complexity and chaos has questioned the naivety that science depends on patterns by establishing a link between determinism and predictability (Nowotny et al. 2001; Sardar and Ravetz 1994). Therefore, it is important that analyses of complex systems are not left solely to scientists, since these systems are transdisciplinary, involving human agents, science, and society (Nowotny 2005).

Mileti (1999) was not alone in recognizing that systems are firmly entrenched in thinking and research on hazards and disasters. According to Gillespie et al. (2004), knowing how to mitigate the negative consequences of natural disasters and respond effectively requires three steps: “understanding the physical and social systems involved in disasters, communicating that understanding clearly to decision-makers, and knowing what interventions may be effective” (Gillespie et al. 2004, p. 82). Complex systems theory provides a holistic approach to integrate these three steps and

understand how complex interactions generate certain behavior, although it is difficult to monitor these complex interactions. Theoretically, the framing of EWS has evolved through systems thinking throughout the last 50 years with growing recognition of the social systems involved in an EWS.

Early Warning Systems within Disaster Management

Individuals, who developed theories on how disasters and EWS operate, as outlined above, were not alone in recognizing that social systems have a significant role in disaster management. In the last century, human geographers have influenced disaster management thinking by challenging top-down expert-driven approaches, by instead suggesting bottom-up locally integrated ones. In the 1930s and 1940s, the “dominant approach” was widely accepted (Wisner 2004), stating that factors such as “material wealth, experience of hazardous events, systems of belief, and psychological considerations are all important in controlling how individuals, social groups, and indeed, whole societies respond to disasters”

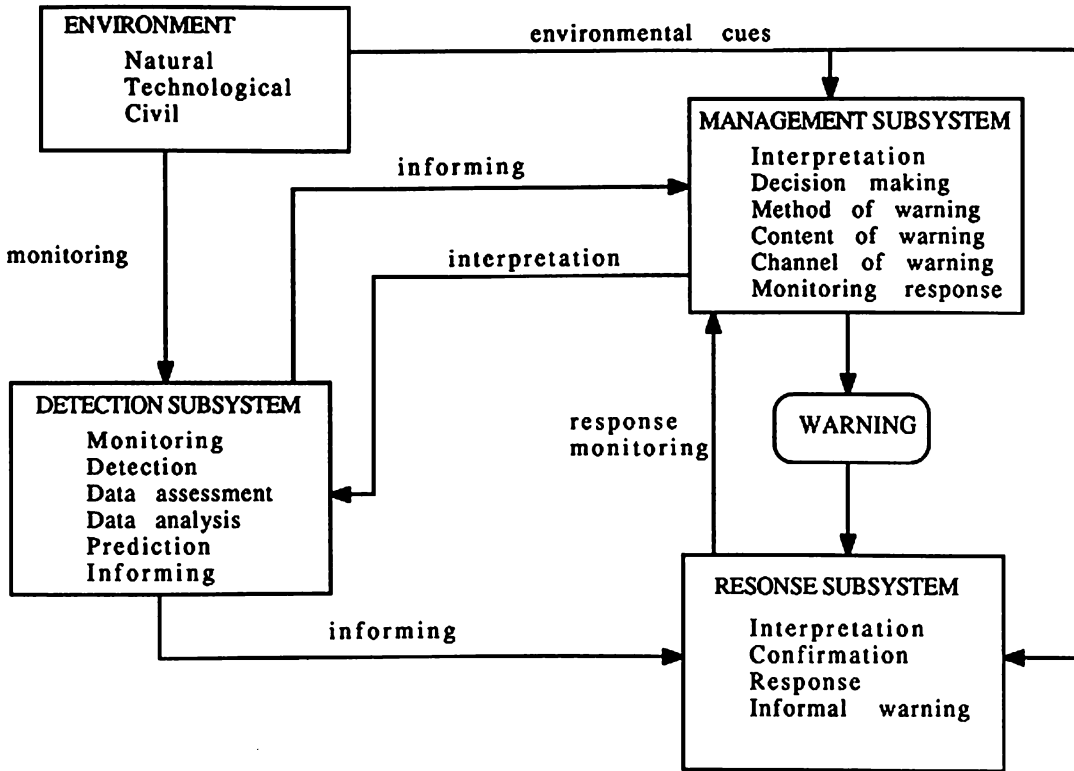
(Chester et al. 2005, p. 416). This approach implied there are adjustments that individuals and societies can make to deal with natural hazards. In the 1980s, Kenneth Hewitt (1983) discussed the inherent complexities in natural-disaster planning in “Interpretations of Calamity” disputing the dominant approach. This new approach adopted the view that most disasters in developing countries are the result of poverty and deprivation rather than extreme natural hazard events, and so those economically or geographically marginalized suffer the most (Susman et al. 1983). Radical alternatives changed the way natural hazards are studied by scientists, social scientists, and policy-makers to emphasize the uniqueness of the location and the socioeconomic and cultural conditions of the society. To be successful, adjustments to hazards should be sensitive to the local environment and be intercultural. The common acceptance of this alternative way of thinking did not occur until the World Conference on Natural Disaster Reduction in 1994, where the published Yokohama Strategy reviewed, in part, how we could transform society to reduce disasters using radical alternatives (UN ISDR 2004).

By the 1990s, EWS became an area of focused research within disaster-management studies. EWS are difficult to understand because they encompass the physical hazard and the context of the “society” affected. Mileti and Sorenson (1990) provide one of the first detailed reviews of EWS from a social science perspective. Based on 200 studies in the USA, they established three key findings. First, variation in the nature and content of warnings has a large impact on whether the public responds. Second, the characteristics of the population receiving the warning affect the response (i.e., gender, ethnicity, and age, and other social, psychological, and knowledge characteristics). Third, many current myths about public response to emergency warnings are at odds with field investigation results, for example, “cry-wolf” syndrome, public panic, and hysteria. These results indicate there is a difference between “ideal” models and those in practice. Drawing on case studies, the authors outline guidance for what information warning messages should contain: the hazard, location, guidance, time, and

sources. For many hazards, including volcanoes, this is extremely challenging to achieve since hazards have different levels of predictability, detectability, certainty, lead time, duration of impact, and visibility as scientific capabilities remain limited, making it difficult to generate “specificity, consistency, accuracy, certainty and clarity” (Mileti and Sorenson 1990, pp. 3–11) in warnings. Mileti and Sorenson (1990) state it is not possible to review EWS in a comprehensive manner by just isolating the social and physical elements, because there is a need to establish organizational effectiveness, work with other organizations, and maintain flexibility during warnings. The report presents a model of EWS (see Fig. 3) with a detection component (monitoring and detection, data assessment and analysis, prediction and informing), emergency management component (interpretation, decision to warn, method and content of warning, and monitoring of response), and response component (interpretation and response). This builds on the White and Haas (1975) model (Fig. 1) by emphasizing the different subsystems and their relationships and institutional roles, rather than only focusing on the relationship between the hazard, warning, and response.

Every aspect of an EWS involves a decision, from interpreting monitoring information, to issuing a warning, to evacuating a town, to the vulnerable individual deciding what to do. Decision-making is still considered under a systemic approach, using a logical sequence between the definitions of the problem, the risk assessment, and its solution, rather than considering the complexities involved (UNDRO 1990). A wide range of institutions have to make decisions about appropriate actions, and the conventional view is these move along a linear chain as shown in Fig. 4, taking a top-down approach. Some countries, such as the USA, adopt this top-down approach to decision-making wherein populations turn to their local civil authorities for information and advice to make informed decisions. Countries that adopt a bottom-up approach place greater responsibility on the individual or community.

In 2004, the Humanitarian Practice Network developed a model of EWS (see Fig. 5) (Twigg



Volcanic Hazards Warnings: Effective Communications of, Fig. 3 The general components of an integrated warning system (Mileti and Sorenson 1990, pp. 2–4)

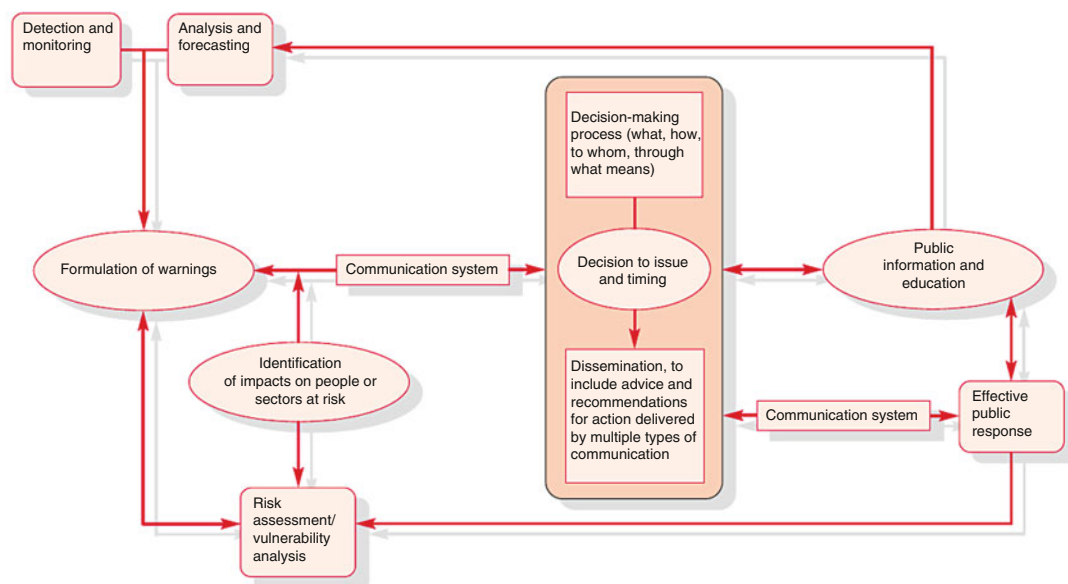


Volcanic Hazards Warnings: Effective Communications of, Fig. 4 The decision-makers within EWS adopting a top-down approach (Fearnley 2011, p. 55)

2004) that shows EWS as a more complex system, with feedback loops and variables, but also identifies the need for risk assessment, understanding vulnerability, and public education. Unlike the linear models shown in Figs. 2, 3, and 4, this model illustrates that decision-making is a core component of EWS and is not a linear process, but the result of feedback from different actors involved in the EWS.

Following the Indian Ocean tsunami of 2004, Hurricane Katrina in the USA in 2005, and the Tohoku tsunami in 2011, recent publications

highlight that EWS are becoming an increasingly topical and important area within disaster risk reduction methods (Glantz 2009; Hall 2007; IFRC 2009). Despite the importance of EWS, research about their application and effectiveness is fragmented, unconsolidated, and patchy. More holistic perspectives of EWS are emerging, although still not widely implemented, viewing them as a system that attempts to interact with a number of complex systems (such as the physical phenomenon, hazard, and society) to provide sufficient warnings for appropriate action to take



Volcanic Hazards Warnings: Effective Communications of, Fig. 5 Generic model of forecasting/warning systems developed by Schlosser, C (Twigg 2004, p. 301)

place. Historically EWS research focused on two key areas: forecasting techniques for natural hazards within the scientific community and exploring strategies to disseminate warnings effectively and credibly to vulnerable populations, often referred to as the “last mile” within disaster studies. Studies on the last mile relate to large literatures on risk perception (Gaillard and Dibben 2008; Slovic 2000), vulnerability (Bankoff et al. 2004; Birkmann 2006; Wisner 2004), resilience (Bankoff 2007; Kelman and Mather 2008), and capacity and communication (Tierney and Dynes 1994). There is a need to also consider the “first mile,” an often overlooked but key component in EWS in an increasingly globalized yet patchily standardized world. This “first mile” relates to the design and operation of EWS and raises questions about how effective they are in communicating warnings and information to all the users of the system. These user groups are growing in diversity as trade and travel becomes increasingly international. Understanding the “first mile” requires investigation into how scientists understand volcanoes, how they manage the associated uncertainties and risks, and how they attempt to manage them both theoretically and practically.

Historically high levels of uncertainty in natural hazard science have resulted in scientists becoming core stakeholders in EWS, due to their expertise and responsibilities, so that EWS became “hazard-focused, linear, top-down, expert/driven systems, with little or no engagement of end-users or their representatives” (Basher 2006, p. 2712). From this, mistrust of expert and local authorities can develop based on criticism that implementing an EWS is a long-term process where local populations can sustain themselves and thus benefit for generations to come. Twigg (2004, p. 306) highlights that:

The bulk of effort and expense is put into transmitting detailed clearly presented information to decision-makers and government emergency management services. Far less effort and funding go into disseminating this information right down to individual communities or households through accessible messages that will warn them and help them to make sensible decisions about how to respond.

To date, there has been little evaluation of the influence of institutional organization and the flow of information between different actors in an EWS on making decisions. Typically, government institutions that manage potential disasters

use simple policy, often prescriptive in manner; however, with the recognition that decision-making is more complex, local practitioners and vulnerable populations are increasingly managing disasters relevant to them using community-based EWS. These EWS are based upon local capabilities and technologies where communities can have ownership, generating an EWS that adopts a bottom-up approach. The idea of community-based EWS has gained momentum (Maskrey 2011), in line with the radical approach developed by Hewitt (1983), and is suggested as an approach to develop people-centric EWS by the UN ISDR PPEW (2006).

Institutional Approaches to EWS

In recent decades, global institutions that provide guidelines and best practices for EWS have increasingly recognized the role of EWS in disaster management, largely the United Nations (UN). The UN General Assembly designated the 1990s as the International Decade for Natural Disaster Reduction (IDNDR), which in 2000 the International Strategy for Disaster Reduction (ISDR) replaced. Throughout the 1990s and 2000s, the UN held a number of EWS conferences resulting in a number of publications (Kuppers and Zschau 2002; UN ISDR 2006a, b). In 2005, the UN established the Hyogo Framework, a global blueprint for disaster-risk reduction (DRR) efforts during the next decade with the goal to substantially reducing disaster losses by 2015. One of its five key priorities for actions is to “identify, assess and monitor disaster risks and enhance early warning” (UN ISDR 2005, p. 6), highlighting growing awareness of the role EWS has within institutional governance.

Following the catastrophic Indian Ocean tsunami of 2004, the Secretary-General of the United Nations called for the development of a global EWS for all natural hazards and communities. It was felt that if an EWS were in place when the tsunami struck the Indian Ocean region, many thousands of lives could have been saved (230,000 are estimated to have been killed in 11 countries (Thieren 2005)). In March 2005, the UN ISDR Platform for the Promotion of Early Warning (PPEW) undertook a global survey to

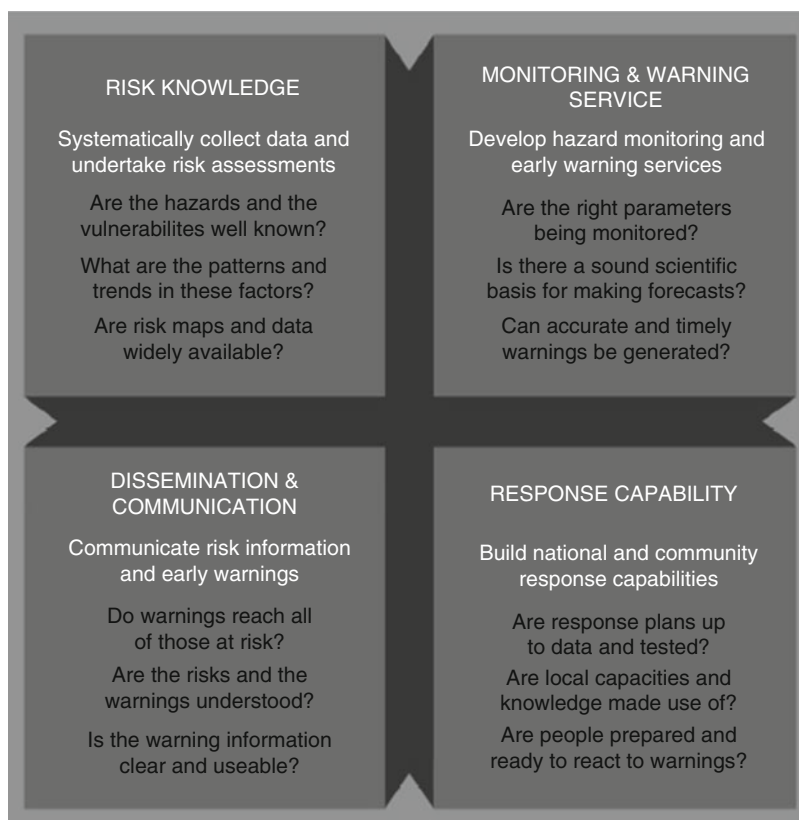
identify existing capacities and gaps in EWS, intended as a wake-up call for governments and other agencies about the value of EWSs in reducing human and economic loss from natural hazards. Published in 2006 the “Global Survey of Early Warning Systems” was the culmination of this research and advocated that EWSs should be “people-centered” (i.e., community based) and encompass spanning four key elements: risk knowledge, monitoring and warning service, dissemination and communication, and response capability (see Fig. 6) (UN ISDR PPEW 2006).

According to the UN, an EWS “can only be effective if the element and the linkages are well-understood, well-designed and well-operated” (Basher 2006, p. 2176). Yet, the model presented in Fig. 6 does not indicate what these linkages are. The survey concludes that the world is far from having the global system for all hazards and communities called for by the UN Secretary-General, but it does make five key recommendations that illustrate the difficulties and contradictions involved in achieving this goal (UN ISDR PPEW 2006, p. vi):

1. Develop a globally comprehensive EWS, rooted in existing EWS and capacities
2. Build national people-centered EWS (i.e., community based)
3. Strengthen the scientific and data foundation for early warnings
4. Fill the main gaps in global early warning capacities
5. Develop the institutional foundations for a global EWS

First, the recommendations raise questions about the viability of uniformity, with different hazards, countries, varying levels of scientific capabilities and communication technologies available, and different local decision-making structures, and institutions involved. Second, it appears contradictory to develop a system that can be globally comprehensive, yet built by the local community. Third, the role of “scientific and data foundation” in preventing EWS failure is questionable, given that this chapter has already reviewed the case study of the Nevado del Ruiz

Volcanic Hazards Warnings: Effective Communications of,
Fig. 6 The elements of a people-centered early warning system (UN ISDR PPEW 2006, p. 2)



tragedy that shows that frequently it is not scientific or technological deficiencies that cause failure, but social and institutional elements. Fourth, identifying gaps may be difficult given that what may be a gap in resources and capabilities for one country may not pose a problem in another, due to differing social and institutional contexts such as available funding. Last, there may be issues with developing institutional foundations for a global EWS when the requirements for emergency response vary in different nations.

Hall (2007) has commented that despite the efforts by the UN events focused on EWSs, there still lacks “coordinated, collaborative international action” (p. 32) to make the move from debate to tangible results. Additionally, Hall outlines that the emphasis within EWS has consequently been more to do with funding of current capabilities and development in science and technology, which has “distracted us from the central issue of address the real needs of the communities and people at risk” (Hall 2007, p. 32). Some

scientists agree, suggesting that they must step outside their “ivory tower” and try to anticipate the consequences of developing warning tools and to make sure they will actually lead to hazard reduction (Malone 2008).

To improve EWS, the UN has called for more effective procedures via standardization and the application of new technologies and enhanced scientific understanding (UN ISDR PPEW 2006). Such a strategy poses two potential weaknesses. First, standardization, by definition, tends to exclude the importance of incorporating local factors into a global procedure (discussed below in section “General Information Statements”). Second, the focus on science and technology implicitly assumes that social and cultural variations are secondary factors, when the materials presented in this chapter clearly illustrate the importance of social context in making EWS effective. The UN has not developed approaches to EWS that consider the complexities involved; instead they focus on the need for developing

global platforms and standardizing; as standardization is reductive, this counteracts systematic approaches to managing crisis. Despite this, standardized methods are frequently used to manage hazards or complex situations.

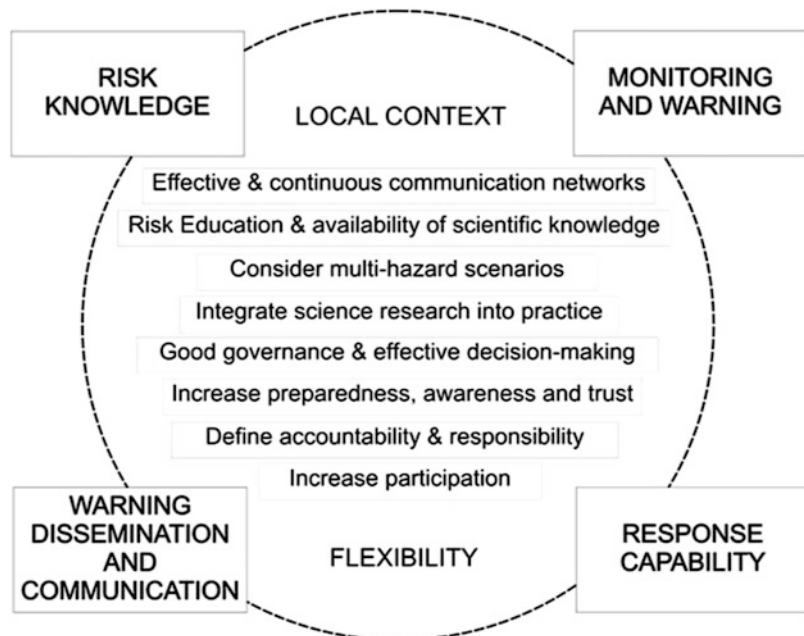
Garcia and Fearnley (2012) highlight that, while an EWS may have four key components as outlined by the UNISDR, it is often the links between these categories that are the focus of systemic failure. Analyzing several case studies conducted over the last 40 years, the authors discovered common emerging factors that improve links between the different components of EWS. They identified four key factors (see also Fig. 7): (1) establishing effective communication networks to integrate scientific research into practice; (2) developing effective decision-making processes that incorporate local contexts by defining accountability and responsibility; (3) acknowledging the importance of risk perception and trust for an effective reaction; and (4) consideration of the differences among technocratic and participatory approaches in EWS, when applied in diverse contexts. In the context of volcano warnings, these vital processes include (1) clear communication (Solana et al. 2008), (2) effective decision-making processes (Leonard et al. 2008), (3) trust building

and participatory activities (Haynes et al. 2008a), and (4) defining accountability and responsibility so people know what to do clearly (Glantz 2004) as per the points above. All these factors show the importance of flexibility and the consideration of local context in making EWSs effective, whereas increasing levels of standardization within EWSs nationally and globally might challenge the ability to incorporate the required local expertise and circumstances.

Volcano Warnings

There are numerous volcano early warning systems in place globally that have been designed to cope with a wide range of hazards. In 2005, the United States Geological Survey (USGS) devised a ranking system called the US National Volcano Early Warning System (NVEWS) (Ewert et al. 2005) to enable recommendations for varying levels of monitoring (Moran et al. 2008). They ranked 169 volcanoes in the USA in a combined assessment of 15 hazard and 9 exposure factors to generate a threat score. Well-established volcano EWSs have evolved over the last centuries, bringing together different practices, technologies, and

Volcanic Hazards Warnings: Effective Communications of,
Fig. 7 Diagram of Early Warning System (EWS) with factors to improve the linking of subsystems (Garcia and Fearnley 2012, p. 133)



procedures; these are discussed extensively with supporting case studies in Fearnley et al. (2017a) and Fearnley and Beaven (2018). Each warning system adopts its own approach unique to the nature of the volcano and the social context in which it operates. Some are technically driven, others are community-based EWS adopting a bottom-up approach. However, most volcano EWS adopt varying types of procedures and protocols to assist in the management of a volcanic crisis. A volcano EWS can be seen of comprising of four key elements (adapted from Potter et al. 2017, p. 4):

1. **Technical:** This relates to the equipment type, deployment (distribution/location/density), telemetry (radio, wire, internet, etc.), visualization (software packages), and analysis; all receive some level of standardization through manufacturing standards, detection limits, and international scientific best practice.
2. **Analytical tools:** Analysis of monitoring data may be further structured through statistical approaches such as expert elicitation, Bayesian event trees, or Bayesian belief networks.
3. **Warning tools:** Notification may be standardized through message content (e.g., standard messages, terminology, alert level criteria), packaging (e.g., bulletins, alert levels, maps), and delivery channels (e.g., phone, internet,

siren). Some standards lend interoperability, such as a Common Alerting Protocol.

4. **Response:** Decision-making and action by the end-user can be standardized to some extent through communication and education approaches and message content.

Typically, volcano warnings are issued using a diverse range of tools, or products, but of significant note is the integration of the volcano alert level system.

Volcano Alert Level Systems

A Volcano Alert Level System (VALS) is the part of a volcano EWS that relates to the processes occurring before and during the issuance of a volcano warning. The USGS defines a volcano alert level system as a “series of levels that correspond generally to increasing levels of volcanic activity” (Gardner and Guffanti 2006, p. 2). As a volcano becomes increasingly active toward eruption, a higher alert level is issued that offers the public and civil authorities a framework they can use to gauge and coordinate their response to a developing volcanic emergency. VALS are based on a linear design, where the alert level assigned is directly proportional to the volcanic activity (see Fig. 8 for an example). In addition, alert levels carry information from the observatory to those who use it in a univalent (one directional) manner.

Volcanic Hazards Warnings: Effective Communications of, Fig. 8 Aviation color code, now internationally adopted by the International Civil Aviation Organization (Gardner and Guffanti 2006, p. 3)

GREEN	Volcano is in typical background, non-eruptive state or, after a change from a higher level, volcanic activity has ceased and volcano has returned to non-eruptive background state.
YELLOW	Volcano is exhibiting signs of elevated unrest above known background level or, after a change from a higher level, volcanic activity has decreased significantly but continues to be closely monitored for possible renewed increase
ORANGE	Volcano is exhibiting heightened or escalating unrest with increased potential of eruption, timeframe uncertain OR eruption is underway with no or minor volcanic-ash emissions [ash-plume height specified, if possible]
RED	Eruption is imminent with significant emission of volcanic ash into the atmosphere likely OR eruption is underway or suspected with significant emission of volcanic ash into the atmosphere [ash-plume height specified, if possible].

Globally, many VALS (also referred to as status levels, condition levels, or color codes) are used providing volcanic warnings and emergency information in relation to volcanic unrest and eruptive activity based on data analysis or forecasts (Potter et al. 2017).

There is a growing body of knowledge that discusses and challenges the role of VALS including: a review of VALS and the role of communication during volcanic crisis (Fearnley and Beaven 2018), assigning an alert level (Fearnley 2013; Winson et al. 2014), standardization of VALS (Fearnley et al. 2012; Potter et al. 2014), and the use and value of alert levels (Papale 2017).

Other Volcano Warning Tools

Supplementing the VALS is a range of other warning tools that can be classified as the following:

Event-Driven (Urgent) Messages

These messages are designed specifically to fulfill users' requirements, for example, by using a number of standardized templates such as products specifically aimed at the aviation sector, such as a "Volcano Observatory Notice for Aviation" (VONA), A Notice to Airmen (NOTAM), A Volcanic Ash Advisory (VAA), and for ground hazard focused users a "Volcano Alert Notification" (VAN). Providing information in specified formats, these are bespoke tools that enable fast and quick communication of the facts at the time of issuance to specific user groups.

Time-Driven (Scheduled) Status Messages

A majority of communication that occurs during a crisis is multivalent, involving communications with several people, usually formalized via a number of protocols such as telephone call-down lists, and meetings between the relevant actors usually as part of a coordination plan, media talking points, and personal communication between the decision-makers. These usually follow pre-arranged schedules and protocols.

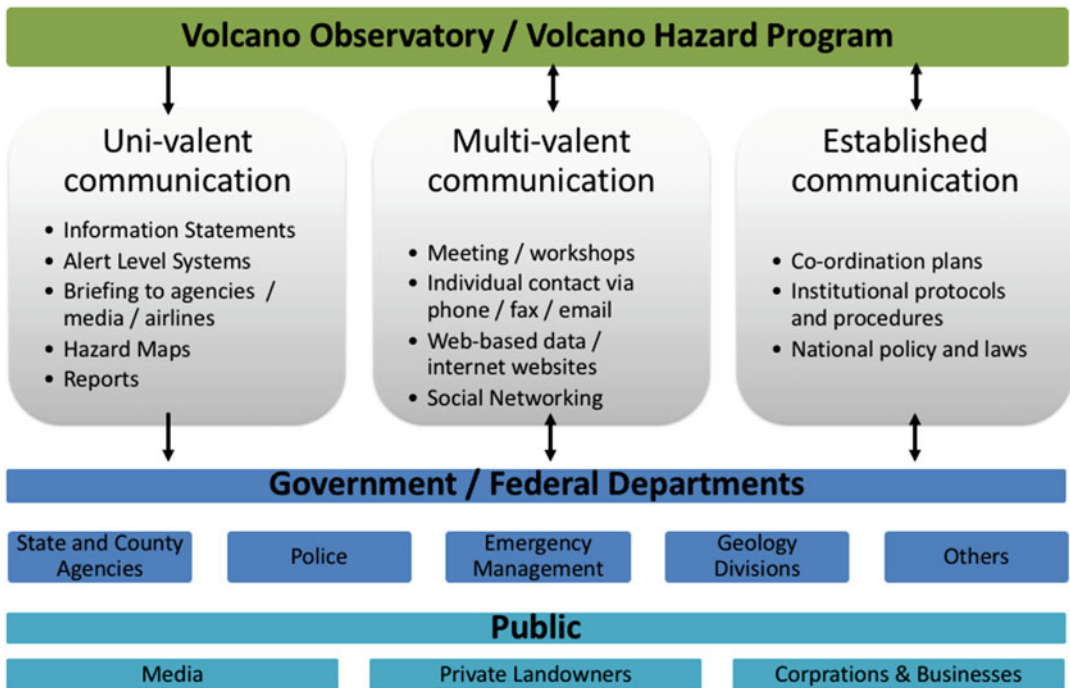
General Information Statements

These are more traditional communication tools that typically consist of information statements, status reports, updates, and other longer text

documents that aim to disseminate the scientific data, interpretation, forecasts, and in some cases guidance. Often, hazard maps and longer reports are also used to release useful information, alongside links to various online resources. For the scientists, information statements provide a greater level of flexibility in communicating information than just issuing an alert level, although they follow a univalent format of information. Scientists tailor these messages to be relevant and of interest to the local users; however, these messages are still limited to text so there is no opportunity for dialogue or for users to add context.

With increasing use of social media and online website and networks, warning information is traveling quicker and to a wider audience than ever, placing pressures on the need for information to be credible, accurate, and of relevance. Most volcano observatories now have their own Twitter or Facebook accounts that support their websites to quickly disseminate messages, alongside the use of Short Message Service (SMS) (Sennert et al. 2015).

While VALS are intended to be linear, in practice the process of issuing a volcanic warning is complex since there is multiplicity of legitimate perspectives, nonlinearity, self-organization, multiplicity of scales, and areas of continuing uncertainty. Therefore, the self-organizing and adaptability of the communication networks provide the flexibility to accommodate the user's needs, requirements, and capabilities in making decisions, and communicating such information to their users. This is not achieved solely through the VALS, which acts more as a "heads up" about the current status of a volcano (although this itself depends on the design of the system that varies by country). During a crisis, the communication that occurs becomes a complex network, or system, that generates feedback loops and enables the communication between the different actors to adapt and evolve as per the requirements of each of the actors involved (see Fig. 9). For every crisis, this system will be different, even if it is the same volcano, because the actors involved and the circumstances are constantly changing. By having a flexible and adaptive communications network, it is possible to accommodate the needs



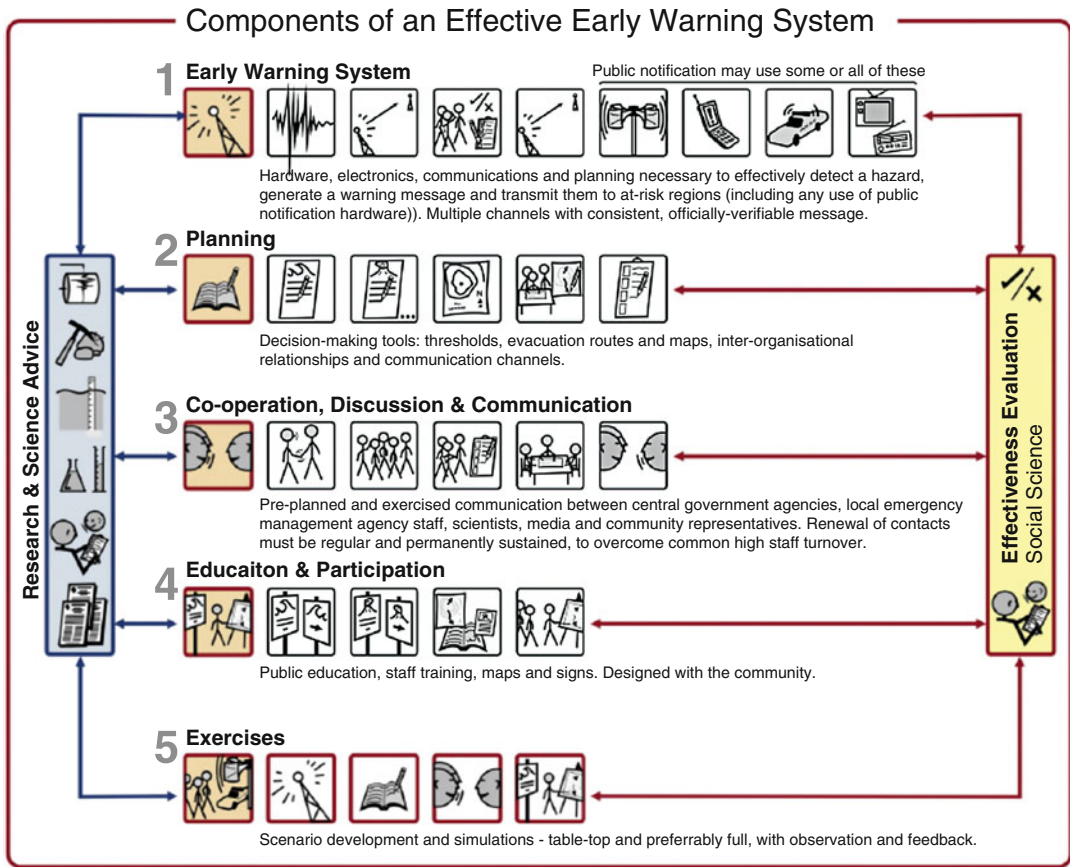
Volcanic Hazards Warnings: Effective Communications of, Fig. 9 Communication tools employed between volcano observatories and key user groups during volcanic crises. (Adapted from Fearnley 2013, p. 1896)

of the diverse range of users and varying hazards over time.

The Challenges for Effective Volcano Warning

Warning effectiveness is not just a function of good hazard knowledge and the generation of a warning message, but needs to be complemented by accurate knowledge of risk and risk management actions (Leonard et al. 2008) (see Fig. 10). However, the effectiveness of an integrated response can be compromised by communication, coordination, training, and organizational constraints (Paton et al. 1998). Once a decision to warn has been made, communication of it in an understandable format to decision-makers and the public is fundamental. It is imperative that all warning communication must be one consistent message, with no contradiction to generate confusion, to help establish trust between the public and other users that the information is correct and useful (Mileti and Sorenson 1990). This creates a problem because often there is scientific controversy.

A number of volcanic crises have highlighted the importance of effective communications between different actors of a volcano EWS that are discussed extensively in Fearnley et al. 2017a but can be summarized as: (1) the need to communicate danger to the public and decision-makers as successfully illustrated by the 1991 eruption of Mt. Pinatubo in the Philippines and the use of the Krafts' educational video (Tayag et al. 1996; Newhall and Solidum 2017); (2) communicating with stakeholders (e.g., civil agencies, land owners, the public) to prepare for volcanic crises as seen at Vesuvius volcano observatory, Italy (Solana et al. 2008); (3) miscommunication between scientists and the media as seen at Soufriere Hills at Guadeloupe during 1976 (Fiske 1984); (4) scaremongering as seen at Galeras volcano in Colombia when it reawakened in 1989 resulting in the public loss of confidence in the scientists (Cardona 1997; Velasco 2000); and (5) within a volcano observatory, culture can shape the ability to communicate and discuss contentious views at a time of crisis as seen at Long



Volcanic Hazards Warnings: Effective Communications of, Fig. 10 Effective early warning systems model from Leonard et al. (2008, p. 204)

Valley Caldera in the USA (Hill 2002; Hill et al. 2017).

Local context is very important to the success of volcano EWS, in particular there are four key local contingencies. First, the political context, as seen in the 1902 eruption of Mt. Pelee that destroyed Saint-Pierre, Martinique, in part the result of politicians who, in the middle of an election “obliging them [the inhabitants] to stay in the city and vote,” effectively, resulting in the death of approximately 30,000 people (Scarth 2002, p. 43). Second, issues of expertise, trust, and credibility as seen in Montserrat (1995–ongoing to present). The decision to evacuate nearly two thirds of the island took much longer than expected, because the government remained uncertain as to the status of the volcano due to challenges over the “experts” view for scientific

advice (Aspinall et al. 2003; Haynes 2008). The possibility of lawsuits and false alarms leading to a loss in credibility and potential inappropriate decisions in the future are consistent concerns (Denis 1995), as also seen during the L’Aquila trial (Alexander 2014; Bretton et al. 2015). Insights from volcano EWS case studies demonstrate there is no one formula for transmitting scientific knowledge, so that the credibility of experts is in a sense always being negotiated and evaluated; therefore, trust cannot be routinized (Wynne 1996). The relationship between scientific expertise and the public is therefore far more complex than typically recognized in calls for “public understanding” that emanate from the scientific establishment. A study in Montserrat discovered the most trusted source for volcanic information is “friends and relatives” (Haynes

et al. 2008a, b), thus highlighting the need for volcanologists to negotiate acceptable levels of risk and trade-offs with the public. Third, the resources available to operate the volcano EWS such as adequate resources and scientists/staff maintain full communications with stakeholders, develop policy, and manage crises. However, this optimal response on an active volcano rarely coincides with an actual crisis (Peterson and Tilling 1993). There are too few observatories, many with limited staff, funding, and equipment for monitoring, resulting in poor communication with local civil officials; moreover, scientists sometimes are so engrossed in their work that they regard interactions with the press and public as annoyances and distractions. The final key local contingency is deterined by the type of volcanic activity, both past and future. Vulnerable populations vary around a volcano, from those that live on the volcano and nearby, to those that live 10's of kilometers away in the river valleys formed by the volcano, to those that live 100's of kilometers away that can be affected by ash. These vulnerable groups have different needs relating to the different hazards and their knowledge of them, and the ability to communicate warning information. Peterson and Tilling (1993, p. 340) identified key factors that lead to some of the complexities involved in operating a volcano EWS. For example, "small, frequent eruptions induce good communications and promote good relations between scientists and the public," as "uncertainty about the outcome of volcanic unrest, especially if major violence is among the possibilities, seems to induce poor inter-relations." In addition, there is recognition that "the public often has unrealistic expectations of scientists' forecasting ability" (p. 348). Therefore, the volcano's eruptive style, activity, and hazards are integral to making a warning relevant to the affected community.

Many lessons have been learnt from volcano EWS including members of the volcanological community who have subsequently reviewed their professional conduct during volcano crises (Newhall et al. 1999). The few case studies addressed demonstrate the value that a more comprehensive understanding of decision-making, communication, and the relevance of local

contexts can make toward designing more effective volcano EWS (Ronan et al. 2000). It is clear volcano EWS need to be flexible in their design to accommodate variation in both the physical hazard and the social context, both of which are locally dependent. This raises questions about the ability for a standardized volcano EWS to achieve its objective.

The Emergence and Challenge of Standardization

Over the last 40 years, volcanic crises have supported the argument that EWS are not linear, but have to negotiate numerous complex systems. This requires a bottom-up approach that considers local context (contingency) that needs to respond to changes over time, and is socially constructed and adapted by the relevant society's requirements. This contradicts increasing levels of standardization in EWS that do not facilitate local flexibility or recognize the complexities involved, and how to best govern them. However, examples outside of disaster management have shown that although standardization is reductive, it helps establish responsibilities and cooperation between the different groups involved. This section draws on examples to review issues that standardization raises that may be relevant to standardizing EWS, addressing the key question as to how a warning can be standardized to consider local context and yet appeal to a diverse range of users?

Why Standardize?

Globally, the levels of standardization in protocols and procedures for disasters and emergency management have risen, including the development of an Indian Ocean Tsunami EWS following the 2004 Boxing Day tsunami. Within the USA, the 9/11 terrorist attacks led to significant changes in government policy resulting in the standardized National Incident Management System (NIMS), the Homeland Security Alert Level, and other alerting and warning protocols for electronic technological warning capabilities. Standardizing warnings is not a new concept, but as disaster practitioners learn more about the complexity of

natural disasters, concerns are being raised that it is increasingly difficult to use “nonlinear” methods of communication and that “faced with the nature and complexity of challenges involved in societal responses to hurricanes [or other hazards], interdisciplinary work that, for example, integrates appropriate meteorological and social science research will be critical” (Gladwin et al. 2009, p. 4). In addition, there appears to be insufficient literature on the effectiveness of standardization as a tool to manage complex disaster-related issues; subsequently, there is minimal understanding of what benefits or limitations standardization can bring. Within disaster studies, guidelines and models for applying standards have been developed, such as consistency and quality control, for developing and using emergency plans (Alexander 2005). Alexander argued that while viewing standards as unnecessarily restrictive and overly prescriptive, they could also help guarantee the quality, content, and relevance of these plans. Given the lack of other data around the standardization of EWS, reviewing other standardized processes such as medical procedures or technological processes can demonstrate issues that standardization raises as a method of managing complexity.

While some regard standardization as a constraint, a number of features also make standardization attractive. First, it improves the “doability” of work to enable scientists to “constrain work practices and define, describe, and contain representations of nature and reality,” and enables a “dynamic interface to translate interest between social worlds” (Fujimura 1987, p. 205). Second, it permits simpler procedures for people to learn from and carry out. Third, in a number of spheres, particularly medical and ethical, it provides answers to concerns relating to the processes or procedures by the public (Hogle 1995). Medical practices regard standardization as necessary to control processes and make outcomes more effective and reproducible. Fourth, standardization provides political ordering and control. In summary, standardization offers a tool to communicate in compatible ways (via language or protocols), ensure minimum quality, and provide a reference point (David and Greenstein 1990).

Standardizing a process is difficult, predominantly because it fixes the process in an ever-changing and dynamic world. In addition, there is no guarantee that researchers or users in different locations will use them in the same way. Scientists tend to “tinker” with standard procedures, often making assumptions of the standard application, so although standardization can increase “doability” it does not guarantee reproducibility (Fujimura 1987). In fact, it can create problems that begin to work against the benefit of standardization, creating tension between the efforts to rationalize work, and changes in the local conditions, which affect the work (Fujimura 1987). Often local practice can render a process less standard, rather than more predictable and uniform (Hogle 1995). Since the cultural, organizational, and institutional relations that characterize a process change, it seems difficult to remove contingency and national variation; for example, medics acting within a standard process bring their own experience and technical contingencies that mean local cultural meanings and categories remain. It is difficult for standardized technologies to be flexible, unless black boxed like a computer, because once a standardized system is in place, it already has a number of users geographically and organizationally that are difficult to change (Hanseth et al. 1996).

Standardization requires establishing boundaries, often in complex scenarios, making it difficult to decide what to leave outside of standardization, and what to include. Most studies on standardization across different practices have demonstrated it is not possible to factor in uncertainty or ignorance when designing a standard, and that knowledge, practices, and technologies of the present shape the standardization. Clearly, these aspects are not static, but to reflect this within the tool of standardization is not easy. Standardizing an EWS that is diverse and pluralistic is particularly challenging.

A Case of Standardization at the United States Geological Survey (USGS) VALS

The emergence and implications of standardization for managing the scientific complexities and diverse agencies involved in volcano crises have

been charted for the USGS, which manages five volcano observatories in Alaska (AVO), Cascades (CVO), Hawaii (HVO), Long Valley (LVO)(now California), and Yellowstone (YVO) (Fearnley 2011). With a wide diversity of volcanic types, many active volcanoes, excellent monitoring resources, and international experience via their Volcano Disaster Assistance Program (VDAP), the USGS provided an excellent case study to review the impact of standardizing their VALS. In 2006, the USGS replaced all previous, locally developed systems at four of its observatories with a common standard. Their VALS comprises of two systems, one for ground hazards, and the other for ash hazards (see Fig. 8), which this case study focuses on. The rationale for standardizing the VALS for ash hazards stemmed from demand for the aviation sector for a standard warning for ash throughout the USA to prevent confusion. Additionally, following the 9/11 terrorist attacks in 2001, increasing levels of standardization for warning procedures and protocols were enforced. There was also a desire to develop uniform warnings that would be more consistent across the USGS.

Research at all five observatories revealed numerous different interpretations in the meaning of an alert level. One such example recalled at both AVO and at HVO concerned a commercial Alaskan pilot flying from Alaska to Hawaii. The pilot, used to flying in Alaska and dealing with the aviation color code frequently in place there, was concerned that the Kilauea volcano on the island of Hawaii was assigned an Orange alert level. Based on his experience with volcanoes in Alaska, he anticipated that the volcano would be exhibiting unrest with increased potential for eruption with ash. When the pilot arrived in Hawaiian airspace, he expected some form of diversion or information (such as a Volcanic Ash Advisory) regarding Kilauea but received nothing and landed with no problems. He later discovered that Kilauea is erupting, but only emitting a small ash plume that prohibits low-level flying within close proximity of the volcano. What he expected was based on his experience with volcanoes assigned alert level Orange in Alaska. Although alert level terms are standardized throughout the

USA, they mean different things to users, in different locations, demonstrating both flexibility and inconsistency in the meaning and interpretation of VALS by users. Often these interpretations build on an individual's local experiences and interactions with a VALS. In addition, the meanings of alert levels change between government agencies. An Orange alert level does not affect the local Volcanic Ash Advisory Centers (VAAC) or the National Weather Service (NWS) in Hawaii as it does in Alaska.

Fearnley (2011) discovered there are tensions between using local VALS, and those that are standardized as summarized in Fig. 11, illustrating that there are benefits associated with both local and national systems. Using a local system provides greater flexibility to adapt to the local needs and integrate the VALS into the management processes of the crisis. However, local systems are becoming increasingly limited by nationally standardized disaster protocols such as the USA National Incident Management System (NIMS). Dependence on common terminology for each alert level may help streamline communications but equally can be misleading as a standardized VALS cannot provide specific information that a locally developed VALS can. Limitations in the ability to provide diversity and pluralism suggest that there may not be enough flexibility in the design. The principle of "one size fits all" does not apply to VALS; they need to adapt to reflect changes in volcanic behavior and their impact on people, and this is better done when they are viewed from a holistic perspective to incorporate all the variables involved, many of which will be unknown prior to the crisis.

The case study concludes that it is difficult for a VALS to be standardized, yet maintain the benefits of a local system and, in addition, to be understood by users both local and global (e.g., aviators). This creates a problem as the more flexible a system becomes, the less standardized it is. Although consistency is frequently identified as a key element of standardization, in practice it does not seem to work. Currently, the USGS standardized VALS works around limitations in flexibility through the many communication and information products and networks developed

Issues	Local (individual USGS Observatories)	National (new standardised system)
Management	Local stakeholders develop close relationships	Streamlines communication within federal agencies reducing confusion
Decision Making	Gear decision on local needs, circumstances and knowledge	Descriptions provide guidelines / criteria, but implications may vary
Communication Methods	Local interpretation likely to be more effective	Common terminology and understanding, but must be known
Users needs	Provides flexibility to local community but global users may be confused	Limits flexibility possible, but two systems specific for their users

Volcanic Hazards Warnings: Effective Communications of, Fig. 11 This graphic compares pros and cons of local (left) and standardized (right) VALS (Fearnley 2011 p. 247)

between the scientists and the users as seen in Fig. 9 (Fearnley and Beaven 2018).

VALS attempt to manage the many complex systems within a volcanic crisis and standardization provides a “one-size-fits-all” approach that is reductive in nature and potentially unable to accommodate local flexibility required to effectively prevent loss of life and minimize economic impact. Instead, standardization appears to be a tool that helps simplify the organizational elements of VALS for policy-makers and large-scale decision-makers, such as government institutions. There is a growing literature reviewing the impacts of the standardization of volcano EWS or VALS from either a national or a local perspective (Potter et al. 2014), although it remains unknown as to how standardization can compromise the need to effectively manage complexity. This is an important issue that warrants further research, and if left answered could generate a number of problems that could lead to further disasters, rather than reducing them.

Future Directions

There is a growing recognition that volcano EWS interface with complex systems and have to negotiate many issues, in addition to globalization, pluralization, and an erosion of expertise. Linear models of EWS are unable to represent the relationships and feedback within these complex

systems because of their constraints in design. If society wants to be prepared for volcano crises then it needs a “truly complex-systemic approach to both the practice and method of science” (Gallopin et al. 2001, p. 223). There is a need to understand the connectedness, relationships, and contexts of volcanic warnings and their dynamics in order to investigate “how the different components and processes interact functionally to generate system responses and emergent properties, how the system adapts and transforms itself” (Gallopin et al. 2001, p. 223). A complex system can be defined as “a system in which large networks of components with no central control and simple rules of operation give rise to complex collective behaviour, sophisticated information processing and adaptation via learning or evolution” (Mitchell 2009, p. 13). In addition, complex systems “exhibit nontrivial emergent and self-organising behaviours” (Mitchell 2009, p. 13).

Complexity theory has been applied to core topics of sociology, spurring a diverse range of studies (Nowotny 2005; Urry 2005b), but this application has not been without contention (Hemaspaandra and Ogihara 2001). Pragmatists suggest that complexity provides a lens that helps us to look at our world and shape our actions, but it should not be seen as the only way to look and do things. Many, however, still have concerns with the reliance of complexity thinking as a method of solving all apparent woes, including volcanic hazards. The most robust critique is that

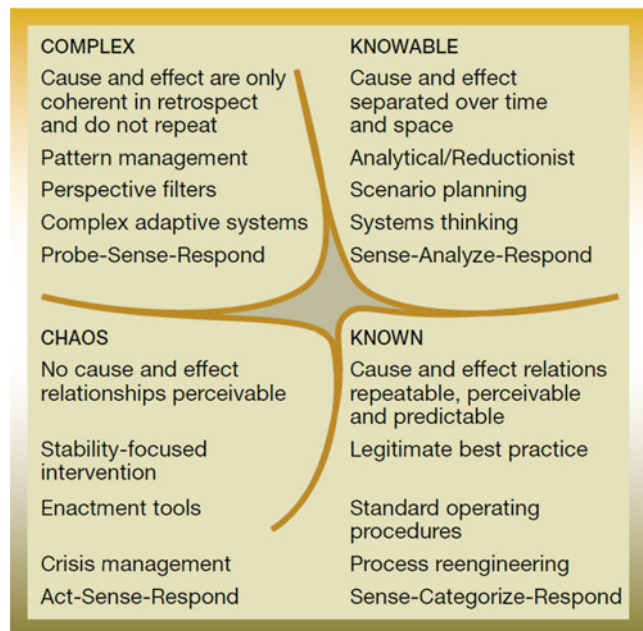
key concepts of complexity are often poorly understood, with issues of their relevance and applicability often ignored or glossed over (Piepers 2006). Regardless, complexity is becoming a popular method to view an ever increasingly interconnected and uncertain world that is unable to be viewed or understood using reductionist methods. There will, however, always be concern that the theory that tries to explain everything may in fact explain nothing at all.

A core element of complexity is that traditional “boundaries” no longer exist, although they may be in place institutionally. Drawing boundaries is an eminently social process and boundaries are routinely drawn between science and nonscience, experts and lay persons, science and politics, and the social and the natural (Gieryn 1983), with consequences for what is taken into account when understanding and managing risk. Complexity appeals most to those who feel that top-down and reductionist approaches are inappropriate in real-world situations because complexity approaches “use rules which promote and permit complex, diverse, and locally fitting behaviour; decentralise, minimise controls and enable local appraisal, analysis, planning and adaption for local fit in different ways” (Chambers 1997, p. 221).

Reconceptualizing Volcano Warnings

Volcano EWS need to manage and interact with a number of complex systems, including the volcano, society, and environment. But how are these complexities negotiated? Historically cost-benefit models and producing policy that tends to revert to a “reductionist” nature have been adopted. The complexity literature offers little resolution as to how to model and use complexity to help manage knowledge and make decisions. There is, however, one such model, Cyneform (Kurtz and Snowden 2003; Snowden 2005), that addresses differing levels of complexity, not by narrowing opportunities through compartmentalizing them into frameworks, but by moving from different stages of known, knowable, complex, and chaotic systems (Fig. 12). Cyneform is a model or approach to policy formation and operational decision-making that recognizes the value of uncertainties and risk, by reducing pattern entrainment. It is a sense-making framework and its value is not in the logical arguments of empirical verification, but in decision-making and facilitating shared understandings to emerge through the many discourses of the decision-makers (Ravetz 1999). This model enables people to make sense of complexity by relaxing three basic assumptions

Volcanic Hazards Warnings: Effective Communications of,
Fig. 12 Cyneform model (Kurtz and Snowden 2003, p. 468)



prevalent in organizational decision-making: assumptions of order, rational choice, and of intent.

The four different domains in the model represent the dynamics of situations, decisions, perspectives, and conflicts when making a decision under uncertain conditions. The boundaries shown are more like phase changes than physical boundaries, so it is possible to consider the problem as it moves between different phases, such as “knowable” to “complex.” This model helps understanding and the interpretation of problems by indicating they are not always static, which is highly applicable to volcanic crises where complexity and chaos are all involved at different stages and within different systems.

By the 2000s, numerous papers highlighted the need for holistic systemic approaches that accommodate the complexities involved, and provide integrated approaches to disaster management (Geis 2000; McEntire and Fuller 2002; Paraskevas 2006). Yet, to date no single overarching theory has been ascribed that captures every variable and issue associated with disasters (McEntire 2004). For this reason, complexity and chaos theories have gained recognition with the growing understanding that disaster responses should be flexible and adaptive (Koehler 1995; Mileti 1999). Likewise, climate change debates have generated a wealth of literature relating to uncertainty, risk, and the plural values of society built on the theory of systems through the concept of complexity and chaos.

Studies of organizational crises that adopt a complex science approach demonstrate that a complexity-informed framework can aid the design of response to a crisis by developing a co-evolving system that essentially self-organizes, learns, and adapts to their dynamically changing environment, a complex adaptive system (CAS) (Paraskevas 2006; Zhong and Low 2009). A CAS is defined as “a number of components of agents, that interact with each other according to sets of rules that require them to examine and respond to each other’s behaviour in order to improve their behaviour” (Stacey 1996, p. 10), and evolve (Axelrod and Cohen 1999). Interactions within the system can produce unexpected patterns or behaviors that can have

unexpected effects on other parts of the system creating nonlinear feedback networks. During a crisis, feedback is required to monitor the progress of the crisis response, and this feedback enables a system to self-correct or modify behavior, learning from experience. Crisis response communication systems, such as volcano EWSs, can be viewed as a CAS where agents self-organize and restructure at a local scale.

In 2008, the Overseas Development Institute (ODI) reviewed the applicability of complexity approaches within real-world crises (Ramalingam et al. 2008). The report highlights that complexity science can generate useful insights into managing complex problems, with a more realistic and holistic approach, supporting useful intuitions, actions, and policy. The idea of self-organization indicates that “actors at all levels of a given system need to be empowered to find solutions to problems, challenging the existing dichotomies of ‘top-down’ versus ‘bottom-up’ so often discussed in disaster practice and international aid agencies” (Ramalingam et al. 2008, p. 62). The concepts of complexity challenge the very method in which current governance conducts its work, as outlined in the following quote (Telford et al. 2006, p. 119):

International agencies need to pay as much attention to how they do things, and their capacities to do them, as they do to the content of their policies and programmes [...] sensitivity to context and the flexibility to adapt to evolving realities are essential, instead of applying predetermined strategies and one-size-fits-all solutions.

Complexity theory and models can provide a tool for practitioners, policy-makers, managers, and researchers to reflect collectively on how they are trying to solve problems, by providing better awareness of why disaster or development and humanitarian work is so problematic.

The Need for an Integrated Volcano Early Warning System

Social contexts affect the use and success of VALS far more than previously acknowledged. There are still many scientific uncertainties within volcanology; scientists are continuously developing theories to better understand the origin, processes, and eruptive behavior of volcanoes and the

numerous associated hazards. However, it is not just scientific constraints involved in determining warnings but also constraints from social and institutional contexts. Volcanic hazards pose a significant problem in society because they generally occur on a longer time frame than political terms or human generations and therefore are not normally a priority. This generally results in limited funding and resources for monitoring volcanoes and conducting research on their past behaviors, and limited volcanic hazard awareness. From an institutional perspective, the wide-ranging impact of volcanic hazards tends to result in the involvement of numerous institutions and agencies, and it is often difficult to maintain communication both within and external to each body involved. Increasing levels of bureaucracy and contending stakeholders mean that decisions can be complex and take a long time to make and implement. Decision-making is a highly pressured process, particularly for the scientists in charge and federal agency users who have a legal obligation to respond. To reduce this pressure, emergency response plans are established prior to crisis to aid and generate communication and understanding; but this is not enough.

Managing volcanic crises requires careful consideration and understanding of how to take action in the context of extreme uncertainty and complexity, both scientifically and socially. To do this successfully a volcano EWS should be fully integrated to cover everything from monitoring and detection, to analysis and interpretation of the data, understanding risk, to communicating information to stakeholders, and generating an effective response. This requires planning, cooperation, the execution of drills, education, and discussion, to name a few processes, between all actors so that during a crisis effective decisions can be made quickly. In summary, this is reconceptualizing a volcano EWS as a complex adaptive system of communication networks.

Bibliography

Alexander D (2005) Towards the development of a standard in emergency planning. *Disaster Prev Manag* 14:158–175

- Alexander DE (2014) Communicating earthquake risk to the public: the trial of the “L’Aquila Seven”. *Nat Hazards* 72:1159–1173
- Aspinall WP, Woo G, Voight B, Baxter PJ (2003) Evidence-based volcanology: application to eruption crises. *J Volcanol Geotherm Res* 128:273–285
- Axelrod R, Cohen MD (1999) *Harnessing complexity: organizational implications of a scientific frontier*. Free Press, New York/London
- Bankoff G (2007) Dangers to going it alone: social capital and the origins of community resilience in the Philippines. *Contin Chang* 22:327–355
- Bankoff G, Frerks G, Hilhorst T (2004) *Mapping vulnerability: disasters, development, and people*. Earthscan Publications, London
- Barton AH (1969) *Communities in disaster: a sociological analysis of collective stress situations*, 1st edn. Doubleday, New York
- Basher R (2006) Global early warning systems for natural hazards: systematic and people-centred. *Phil Trans R Soc* 364:2167–2182
- Bertalanffy LV (1975) *Perspectives on general system theory: scientific-philosophical studies*. G. Braziller, New York
- Bertalanffy LV, Woodger JH (1933) [Kritische Theorie der Formbildung.] *Modern theories of development. An introduction to theoretical biology ...* Translated and adapted by JH Woodger. p x. 204. Oxford University Press: London
- Birkmann J (2006) *Measuring vulnerability to natural hazards: towards disaster resilient societies*. United Nations University, Tokyo
- Bretton RJ, Gottsmann J, Aspinall WP, Christie R (2015) Implications of legal scrutiny processes (including the L’Aquila trial and other recent court cases) for future volcanic risk governance. *J Appl Volcanol* 4:18
- Capra F (1996) *The web of life: a new synthesis of mind and matter*. Flamingo, London
- Cardona OD (1997) Management of the volcanic crises of Galeras volcano: social, economic and institutional aspects. *J Volcanol Geotherm Res* 77:313–324
- Chambers R (1997) *Whose reality counts? Putting the first last*. Intermediate Technology, London
- Chester DK, Marti J, Ernst GG (2005) Volcanoes, society, and culture. In: *Volcanoes and the environment*. Cambridge University Press, Cambridge, pp 404–439
- David PA, Greenstein S (1990) The economics of compatibility standards: An introduction to recent research. *Economics of innovation and new technology* 1(1-2):3–41
- Day S, Fearnley C (2015) A classification of mitigation strategies for natural hazards: implications for the understanding of interactions between mitigation strategies. *Nat Hazards* 79:1219–1238
- Denis H (1995) Scientists and disaster management. *Disaster Prev Manag* 4:14–19
- Ewert JW, Guffanti M, Murray TL (2005) *An assessment of volcanic threat and monitoring capabilities in the United States: framework for a National Volcano Early Warning System*. U.S. Geological Survey, Reston
- Fearnley CJ (2011) *Standardising the USGS volcano alert level system: acting in the context of risk, uncertainty*

- and complexity. UCL (University College London), Great Britain
- Fearnley CJ (2013) Assigning a volcano alert level: negotiating uncertainty, risk, and complexity in decision-making processes. *Environ Plan A* 45:1891–1911
- Fearnley CJ, Beaven S (2018) Volcano alert level systems: managing the challenges of effective volcanic crisis communication. *Bull Volcanol* 80(5):46
- Fearnley C, McGuire W, Davies G, Twigg J (2012) Standardisation of the USGS Volcano Alert Level System (VALS): analysis and ramifications. *Bull Volcanol* 74:2023–2036
- Fearnley C, AEG W, Pallister J, Tilling RI (2017a) Volcano crisis communication: challenges and solutions in the 21st century. In: *Observing the volcano world: volcanic crisis communication*. Springer, Cham
- Fearnley CJ, Bird DK, Haynes K, McGuire W, Jolly G (eds) (2017b) *Observing the volcano world: volcanic crisis communication advances in volcanology*. Springer. <https://doi.org/10.1007/978-3-319-44097-2>
- Fiske RS (1984) Volcanologists, journalists, and the concerned local public: a take of two Citites in the Eastern Caribbean. In: *Studies in geophysics: explosive volcanism: inception, evolution, and hazards*. National Academy Press, Washington, DC, pp 170–1976
- Form W, Nostow S (1958) *Community in disaster*. Harper & Row, New York
- Foster HD (1980) *Disaster planning: the preservation of life and property*. Springer, New York
- Fujimura JH (1987) Constructing ‘Do-Able’ problems in cancer research: articulating alignment. *Soc Stud Sci* 17:257–293
- Gaillard JC, Dibben CJL (2008) Volcanic risk perception and beyond. *J Volcanol Geotherm Res* 172:163–169
- Gallopin GC, Funtowicz S, O’Connor M, Ravetz J (2001) Science for the twenty-first century: From social contract to the scientific core. *Int Soc Sci J* 53:219–229
- Garcia C, Fearnley CJ (2012) Evaluating critical links in early warning systems for natural hazards. *Environ Hazards* 11:123–137
- Gardner CA, Guffanti MC (2006) U.S. geological survey’s alert notification system for volcanic activity. U.S. Geological Survey, Reston
- Geis D (2000) By design: the disaster resistant and quality-of-life community. *Nat Hazard Rev* 1:151–160
- Gieryn TF (1983) Boundary-work and the demarcation of science from non-science-strains and interests in professional ideologies of scientists. *Am Sociol Rev* 48:781–795
- Gillespie DF, Perry RW (1976) An integrated systems and emergent norm approach to mass emergencies. *Mass Emergencies* 1(4):303–312
- Gillespie DF, Robards KJ, Cho S (2004) Designing safe systems: Using system dynamics to understand complexity. *Nat Hazards Rev* 5:82–88
- Gladwin H, Lazo JK, Morrow BH, Peacock WG, Wiloughby HE (2009) Social science research needs for the hurricane forecast and warning system. *Bull Am Meteorol Soc* 90:25–29
- Glantz MH (2004) Usable science 8: early warning systems: do’s and don’ts, report of workshop 20–23 Oct 2003, Shanghai
- Glantz MH (2009) *Heads up! Early warning systems for climate, water and weather-related hazards*. United Nations University Press, New York
- Hall ML (1990) Chronology of the principal scientific and governmental actions leading up to the November 13, 1985 eruption of Nevado-Del-Ruiz. Colombia *J Volcanol Geothermal Res* 42:101–115
- Hall PH (2007) Early warning systems: reframing the discussion. *Aust J Emerg Manag* 22:32–36
- Hanseth O, Monteiro E, Hatling M (1996) Developing information infrastructure: the tension between standardization and flexibility science. *Technol Hum Values* 21:407–426
- Haynes K (2008) Volcanic isnad in crises: investigating environmental uncertainty and the complexities it brings. *Aust J Emerg Manag* 21:21–28
- Haynes K, Barclay J, Pidgeon N (2008a) The issue of trust and its influence on risk communication during a volcanic crisis. *Bull Volcanol* 70:605–621
- Haynes K, Barclay J, Pidgeon NF (2008b) Whose reality counts? Factors affecting the perception of volcanic risk. *J Volcanol Geothermal Res* 172:259–272
- Hemaspaandra LA, Ogihara M (2001) *The complexity theory companion*. Springer, New York
- Hewitt K (1983) *Interpretations of calamity from the viewpoint of human ecology*. Allen & Unwin, Boston
- Hill DP, California. Division of Mines and Geology., Geological Survey (U.S.) (2002) *Response plan for volcano hazards in the Long Valley Caldera and Mono Craters region, California* U.S. Geological Survey bulletin 2185
- Hill DP, Mangan MT, McNutt SR (2017) *Volcanic unrest and hazard communication in long valley volcanic region, California*
- Hogle LF (1995) Standardization across non-standard domains: the case of organ procurement science. *Technol Hum Values* 20:482–500
- IFRC (2009) *World disasters report 2009- focus on early warning, early action*. International Federation of Red Cross and Red Crescent Societies, Geneva
- Kelman I, Kelman I (2017) Linking disaster risk reduction, climate change, and the sustainable development goals. *Disaster Prev Manag Int J* 26:254–258
- Kelman I, Mather TA (2008) Living with volcanoes: the sustainable livelihoods approach for volcano-related opportunities. *J Volcanol Geotherm Res* 172:189–198
- Keys H, Green P (2010) Mitigation of volcanic risks at Mt Ruapehu, New Zealand. In: *Proceedings of the mountain risks international conference*. Firenze, pp 485–490
- Kilburn CRJ (2003) Multiscale fracturing as a key to forecasting volcanic eruptions. *J Volcanol Geotherm Res* 125:271–289
- Kim DH (1994) *Systems thinking tools: a user’s reference guide*. Toolbox reprint series. Pegasus Communications, Cambridge
- Koehler GA (1995) What disaster response management can learn from chaos theory: conference proceedings.

- In: Koehler GA (ed) What disaster response management can learn from chaos theory, 2001. California Research Bureau, California State Library, Sacramento, pp 293–308
- Kuppers AN, Zschau J (2002) Early warning systems for natural disaster reduction. In: Early warning systems for natural disaster reduction. Springer, Springer-Verlag Heidelberg, pp 817–821
- Kurtz CF, Snowden DJ (2003) The new dynamics of strategy: sense-making in a complex and complicated world. *IBM Syst J* 42:462–483
- Leonard GS, Johnston DM, Paton D, Christianson A, Becker J, Keys H (2008) Developing effective warning systems: ongoing research at Ruapehu volcano. *N Z J Volcanol Geotherm Res* 172:199–215
- Malone S (2008) A warning about early warning. *Seismol Res Lett* 79:603–604. <https://doi.org/10.1785/gssrl.79.5.603>
- Maskrey A (2011) Revisiting community-based disaster risk management. *Environ Hazards* 10:42–52. <https://doi.org/10.3763/ehaz.2011.0005>
- McEntire DA (2004) The status of emergency management theory: issues, barriers, and recommendations for improved scholarship. In: Paper presented at the FEMA higher education conference, Emmitsburg, 8th June
- McEntire DA, Fuller C (2002) The need for a holistic theoretical approach: an examination from the El Nino disasters in Peru. *Disaster Prev Manag* 11:128–140
- Mileti D (1999) Disasters by design: a reassessment of natural hazards in the United States. Joseph Henry Press, Washington, DC
- Mileti DS, Sorenson JH (1990) Communication of emergency public warnings: a social science perspective and state-of-the-art assessment. Oak Ridge National Laboratory
- Mitchell M (2009) Complexity: a guided tour. Oxford University Press, Oxford/New York
- Moran SC et al (2008) Instrumentation recommendations for volcano monitoring at US volcanoes under the national volcano early warning system. Geological Survey (US)
- Newhall C, Solidum RU (2017) Volcanic hazard communication at Pinatubo from 1991 to 2015
- Newhall C et al (1999) Professional conduct of scientists during volcanic crises. *Bull Volcanol* 60:323–334
- Nowotny H (2005) The increase of complexity and its reduction: emergent interfaces between the natural sciences, humanities and social sciences. *Theory Cult Soc* 22:15–31
- Nowotny H, Scott P, Gibbons M (2001) Re-thinking science: knowledge and the public in an age of uncertainty. Polity Press, Cambridge
- Oppenheimer C (2011) Eruptions that shook the world. Cambridge University Press, Cambridge
- Papale P (2017) Rational volcanic hazard forecasts and the use of volcanic alert levels. *J Appl Volcanol* 6:13
- Paraskevas A (2006) Crisis management or crisis response system? A complexity science approach to organizational crises. *Manag Decis* 44:892–907
- Paton D, Johnston D, Houghton BF (1998) Organisational response to a volcanic eruption. *Disaster Prev Manag* 7:5–13
- Peterson DW, Tilling RI (1993) Interactions between scientists, civil authorities and the public at hazardous volcanoes. In: Kilburn CRJ, Luongo G (eds) Active lavas. UCL Press, London, pp 339–365
- Piepers I (2006) The International System: “At the Edge of Chaos”. arXiv preprint nlin/0611062.
- Potter SH, Jolly GE, Neall VE, Johnston DM, Scott BJ (2014) Communicating the status of volcanic activity: revising New Zealand’s volcanic alert level system. *J Appl Volcanol* 3:13
- Potter SH, Scott BJ, Fearnley CJ, Leonard GS, Gregg CE (2017) Challenges and benefits of standardising early warning systems: a case study of New Zealand’s volcanic alert level system
- Ramalingam B, Jones H, Reba T, Young J (2008) Exploring the science of complexity: ideas and implications for development and humanitarian efforts. In: Working paper 285. Overseas Development Institute (ODI), London
- Ravetz IR (1999) What is post-normal science. *Futures-the Journal of Forecasting Planning and Policy*, 31 (7):647–654
- Ronan KR, Paton D, Johnston DM, Houghton BF (2000) Managing societal uncertainty in volcanic hazards: a multidisciplinary approach. *Disaster Prev Manag* 9:339–349
- Sardar Z, Ravetz JR (1994) Complexity: fad or future? *Futures* 26:563–567
- Scarth A (2002) La catastrophe Mount Pel’ee and the destruction of Saint-Pierre. Terra Publishing, Harpenden
- Senge PM (1990) The fifth discipline: the art and practice of the learning organization, 1st edn. Doubleday/Currency, New York
- Sennert SS, Klemetti EW, Bird DK (2015) Role of social media and networking in volcanic crises and communication
- Sigurdsson H, Houghton B, McNutt S, Rymer H, Stix J (2015) The encyclopedia of volcanoes. Elsevier, San Diego
- Slovic P (2000) The perception of risk. Risk, society, and policy series. Earthscan Pub, London
- Snowden D (2005) Strategy in the context of uncertainty. *Handb Bus Strateg* 6:47–54
- Solana MC, Kilburn CRJ, Rolandi G (2008) Communicating eruption and hazard forecasts on Vesuvius. *South Ital J Volcanol Geotherm Res* 172:308–314
- Sparks RSJ, Aspinall W (2004) Volcanic activity; frontiers and challenges in forecasting, prediction and risk assessment. In: Stephen R, Sparks J (eds) The state of the planet; frontiers and challenges in geophysics. American Geophysical Union, Washington, DC
- Stacey RD (1996) Complexity and creativity in organizations. Berrett-Koehler, San Francisco
- Stirling A (2007) Risk, precaution and science: towards a more constructive policy debate: talking point on the precautionary principle. *EMBO Rep* 8(4):309–315

- Susman P, O'Keefe P, Wisner B (1983) Global disasters, a radical interpretation. In: Hewitt K (ed) *Interpretations of calamity: from the viewpoint of human ecology*. Allen & Unwin, London, pp 263–283
- Swanson DA, Casadevall TJ, Dzurisin D, Holcomb RT, Newhall CG, Malone SD, Weaver CS (1985) Forecasts and predictions of eruptive activity at Mount St. Helens, USA: 1975–1984. *J Geodyn* 3:397–423
- Tayag J, Insauriga S, Ringor A, Belo M, Newhall CG, Punongbayan RS (1996) People's response to eruption warning: the Pinatubo experience, 1991–92. In: *Fire and mud: eruptions and lahars of mount Pinatubo*, Philippines. University of Washington Press, Seattle/London, pp 87–99
- Telford, J, J Cosgrave and R Houghton (2006) Joint Evaluation of the international response to the Indian Ocean tsunami: Synthesis Report. London: Tsunami Evaluation Coalition
- Thieren M (2005) Asian tsunami: death-toll addiction and its downside. *Bull World Health Org* 83:82
- Tierney KJ, Dynes RR (1994) In: Dynes RR, Tierney KJ (eds) *Disasters, collective behavior, and social organization*. University of Delaware Press/Associated University Presses, Newark, London/Cranbury
- Tobin GA, Whiteford LM (2002) Community resilience and volcano hazard: the eruption of Tungurahua and evacuation of the Faldas in Ecuador. *Disasters* 26:28–48
- Twigg J (2004) *Disaster risk reduction: mitigation and preparedness in development and emergency programming*. HPN
- UN ISDR (2004) Yokohama strategy and plan of action for a safer world: guidelines for natural disaster prevention, preparedness and mitigation. In: *World conference on disaster reduction, Yokohama, 23–27 May 1994*
- UN ISDR (2005) *Hyogo framework for action 2005–2015: building the resilience of nations and communities to disasters*
- UN ISDR (2006a) Early warning-from concept to action-the conclusions. In: *Third international conference on early warning: from concept to action, Bonn, 27–29 Mar 2006*
- UN ISDR (2006b) Early warning – from concept to action: the conclusions of the third international conference on early warning (EWC III). In *Bonn, 27–29 Mar 2006*, p 32
- UN ISDR PPEW (2006) *Global survey of early warning systems: an assessment of capacities, gaps and opportunities toward building a comprehensive global early warning system for all natural hazards*
- UNDRO (1990) *Mitigating natural disasters: phenomena, effects and options*. United Nations Geneva
- UNISDR (2007) Terminology. <http://www.unisdr.org/we/inform/terminology-letter-m>. Accessed 27 Feb 2014
- Urry J (2005a) The complexities of the global. *Theory Cult Soc* 22:235–254
- Urry J (2005b) The complexity turn theory. *Cult Soc* 22:1–14
- Velasco MLC (2000) Mt Galeras: activities and lessons to be learned philosophical transactions of the Royal Society of London series. *Math Phys Eng Sci* 358:1607–1617
- Voight B (1990) The 1985 Nevado-Del-Ruiz volcano catastrophe-anatomy and retrospection. *J Volcanol Geotherm Res* 42:151–188
- White D (1995) Application of systems thinking to risk management: a review of the literature. *Manag Decis* 33:35–45
- White GF, Haas JE (1975) *Assessment of research on natural hazards*. MIT Press, Cambridge
- Winson AE, Costa F, Newhall CG, Woo G (2014) An analysis of the issuance of volcanic alert levels during volcanic crises. *J Appl Volcanol* 3:14
- Wisner B (2004) *At risk: natural hazards, people's vulnerability, and disasters*, 2nd edn. Routledge, London/New York
- Wisner B, Gaillard JC, Kelman I (2012) *Handbook of hazards and disaster risk reduction and management*. Routledge, New York
- Wynne B (1996) May the sheep safely graze? A reflexive view of the expert-lay knowledge divide. In: Lash S, Szerszynski B, Wynne B (eds) *Risk, environment and modernity: towards a new ecology*. Sage, London, pp 44–83
- Zhong Y, Low SP (2009) Managing crisis response communication in construction projects – from a complexity perspective. *Disaster Prev Manag* 18:270–282

Index

A

Absolute time, 648, 654
Accretionary lapilli, 584, 586
Accretionary prism, 17, 18
Accretionary wedge, 56
Added-mass effect, 205
ADvanced CIRCulation Model For Oceanic, Coastal And Estuarine Waters, 126
Age estimation methods, tsunami deposits, 145–146
Age-specific eruption rate, 654
Akaike information criterion (AIC), 343, 344, 447, 691
Akaike's Bayesian Information Criterion (ABIC), 45
1964 Alaska earthquake, 4
Aleatory uncertainty(ies), 25, 45, 46
Aleatory variability, 91
1946 Aleutian Islands earthquake, 4
Alpha spectrometry, 145
Amplitude dispersion, 119
Analogue approach, explosive volcanic eruptions
 caldera collapse, 610–612
 cinder cones, 607, 609
 decompression-driven flows, 590–596
 degassing-driven flows, 594, 596–602
 explosion consequences, 606–612
 high VEI events, 588–589
 low VEI events, 602–607
 post-fragmentation flows, 589–590
 pyroclastic flows, 608, 610
Analytical deformation-source models, 522–523
Andesite, 622, 623
Assimilation, 503
Auckland Volcanic Field, 673, 677, 692
Autocorrelation function, 653
Autoregressive (AR) model, 343, 427, 443
Autoregressive-moving average (ARMA) model, 448

B

Basalt, 503
Basaltic magma, 636, 638
Bathymetric lidars, 516
Bayes, Thomas, 92
Bayesian estimation, 680
Bayesian inference, 522, 525
 techniques, 44

Bayesian method(s), 44, 503
Bayesian probabilistic hazard analysis
 data weighting factors, 104
 tsunami forecast, 97
Bayesian statistics, 91
Bayesian Type II maximum likelihood technique, 44
Bayes' theorem, 94
Benchmark, 503
Bárdarbunga volcano, 551–552
Bernoulli probability, 679
Bias, 91
Bingham liquid, 637, 638
Bingham rheology, 637
Binomial model, 108
Body wave magnitude method (bMag), 308
Bore, 187
Bottom friction, 127, 131
Boundary-element method (BEM), 524
Boussinesq models, 127–130
Branching process models, 43
Brine, 503
Broad-band body-wave (mB) magnitude, 299
Bubble dynamics, 471
Bubbly liquids, 471, 487
Buoyant forces, 187, 202, 207
Byerlee's law of rock friction, 57

C

Cahn-Hilliard diffusion-advection equation, 496
Cahn-Hilliard-Navier-Stokes system, 496
Calderas, 503
Campaign-style GPS, 511
14C dating, 146
Central volcano, 539
Centroid Moment Tensor (CMT) solutions, 339, 343, 367–368
Cesium-137 loss and gain identification, 146
Chaotic system, 662, 664
Choked flow, 481, 484
Chouet's three-dimensional model, 476
Cinder cones, 607, 609
Civil society, 384, 385
Closure depth, 135
Clustering, 28

CMT centroid moment tensor, 299
 Coastal armoring, 187
 Coastal tsunami protection structures, 198
 COBRAS, 128
 Common Alerting Protocol, 729
 Completeness, 91
 Complex adaptive system (CAS), 738
 Complexity theory, 736, 738
 Conditional probability, 91
 Conduit, 620, 623–634, 639–642
 flow, 638
 resistance, 636–637
 Conjugacy, 91
 Conservation equations, 496–497
 Continuously-operating GPS (CGPS) stations, 511
 Contourites, 161
 Convolution, 299
 Coriolis force, 351
 Cornell Multi-grid Coupled Tsunami Model (COMCOT), 126
 Cornell University Long and Intermediate Wave Modeling Package, 127
 Correlation function, 662
 Coseismic phase, 248
 Coseismic seafloor deformation (CSD), 64–66
 Coulomb friction criterion, 57
 Coulomb plasticity, 56, 57
 Coulomb stress, 685
 Coulomb wedge, 57, 66
 dynamic, 60–62
 stable and critical, 58–60
 Cox process, 654
 Crack resonance, 476
 Crack wave, 449–452, 478, 492, 493
 Cross-lamination, 135
 Crustal deformation
 definition, 540
 in Iceland, 543–555
 Cryptodome, 503
 Crystal-growth rate, 629
 Crystallization, 622, 624, 627, 629, 631, 633–636, 640–642
 Cylindrical conduits, 623, 627, 628
 Cyneform model, 737

D

Damming effect, 206
 Darcy–Wiesbach type friction factor, 127
 Data acquisition, 521
 Debris impact forces, 187, 204–207
 Deconvolution, 299
 Deep earthquake, 299
 Deep-ocean Assessment and Reporting of Tsunamis (DART), 19, 20, 76, 84, 196, 337, 340, 363
 Deflation, 503
 Deformation, 504
 Deformation source modeling, goal of, 521
 Degassing, 622, 624, 627, 633
 Diffuse-interface model, 494

Digital elevation model (DEM), 213, 504
 Dikes, 504, 539
 Dilatometer, 504
 Dip, 247
Dirichlet distribution, 95
 Disaster risk reduction, 718
 Discrete Fourier transform (DFT), 442, 446
 Dislocation, 504
 Dispersive tsunami, 121
 Distant-source tsunamis, 197
 Distant tsunami, 187
 3D Navier–Stokes equation model, 128
 Dome-building eruptions, 621, 629, 641
 Drag force, 187, 201
 Dry tilt measurements, 539
 Dykes, 623, 625, 635, 638–641
 Dynamic Coulomb wedge, 60–62
 Dynamic earthquake model, 25
 Dynamic earthquake rupture, 29
 Dynamic modeling, 46
 Dynamic models, 47
 Dynamic rupture models, 30–36, 47
 Dynamic rupture propagation models, 49

E

Early warning systems (EWS)
 complex systemic approach, 721
 within disaster management, 722–726
 for disaster-risk reduction (DRR) measures, 719
 economic, political and social circumstances, 719
 general systems theory, 720
 institutional approaches, 726–728
 linear model, 736
 linear processes, 720
 theoretical approaches, 720
 Earthquake(s), 27, 393
 dynamic rupture models, 30–36
 fault, 247
 magnitude, 337, 338
 static slip models, 28–30
 Earthquake early warning (EEW), 366–368
 Earthquake source parameters, 307
 bMag method, 308, 309
 mantle magnitude (M_m) method, 314–315
 moment magnitude M_W , 307
 pMag scale, 308
 P-wave moment magnitude (M_{wp}) method, 309
 W-phase method, 315–317
 Edge wave, 30
 modes, 30
 theory, 30
 Effective stress ratio, 59
 Elapsed time, 654
 Elastic finite-element model, 30
 Elasticity, 504
 Electronic distance-meter (EDM), 504
 Energy magnitude (M_E) scale, 299
 Ensemble Kalman filter technique, 86
 Envelope decay, 37

- Envelope of tsunami, 37, 38
- Epicenter, 247
- Epidemic Type Aftershock Sequence (ETAS) model, 43
- Epistemic uncertainty, 91
- Equilibrium state, 494–496
- Estimation, 91
- Exceedance probability, 91
- Explosive eruption, 621–624, 633, 641
- Explosive volcanic eruptions
 - analogue approach, 588–612
 - caldera collapse, 610–612
 - cinder cones, 607, 609
 - consequences, 571
 - decompression-driven flows, 576, 578–581, 590–596
 - degassing-driven flows, 594, 596–602
 - explosion consequences, 606–612
 - exsolution, 574–577
 - fall velocity of silicate fragments, 583
 - fragment electrification, 583
 - fragment size distributions, 581–583
 - high VEI events, 588–589
 - hydrovolcanism, 572–574
 - low VEI events, 602–607
 - magma rheology, 563–568
 - nucleation and diffusion, 568–570
 - permeability, 570–571
 - post-fragmentation flows, 589–590
 - pyroclastic flows, 586–588, 608, 610
 - silicate fragments and aggregation, 583–588
 - volatile solubility, 568, 569
 - volcanic materials, 563–571
 - volcanic processes, 571–588
- Exponential distribution, 44
- Exposure (elements at risk), 213
- Exsolution, 574–577
- Extended information criterion (EIC), 447
- Extensive CGPS networks, 512
- Extreme value theory, 651
- Extrusive eruptions, 623–625
- Eyjafjallajökull, 552
- F**
- Fault gouge, 57
- Fault model, 337
- Fault parameters, earthquake tsunamis
 - effect of, 74–75
 - heterogeneous fault motion, 82
 - nonlinear inversion methods, 83–84
 - real-time data assimilation, 84–86
 - seafloor deformation, 73–74
 - trial and error approach, 81–82
 - tsunami heights, inversion of, 84
 - uncertainty of, 338
 - waveform inversion, 82–85
- Fault surface, 247
- Finite difference method (FDM), 524
- Finite-element method (FEM), 34, 523
- Fission waves, 124
- Fissure swarm, 539
- Fixed grid method, 129, 130
- Fluid-filled crack model, 452–457
- Fluid-solid interactions
 - crack wave, 449–453
 - fluid-filled crack model, 452–457
- e*-folding decay constant, 37
- e*-folding time, 40, 41
- Footwall, 247
- Forecast point, 351–353, 356, 357
- Form drag, 187, 201
- Forward models, 521
- Fractionation (fractional crystallization), 504
- Fracture mechanics, 18
- Fragmentation bomb, 576, 578
- Fragment size distributions (FSD), 581–583
- Frequency dispersion, 119, 121, 122
- Frequentist statistics, 91
- Friction
 - behavior, 32
 - coefficient, 33
 - law, 31
 - slip-weakening friction, 32
 - time weakening friction, 32
- Fukushima catastrophe
 - body of experience, 374–375
 - cause of, 375
 - collective and adaptive learning process, 373
 - disruptive event, 374–375
 - disruptive knowledge, 377–381
 - high-tech, value-added, or knowledge-based society model, 385–387
 - instability of political life, 375
 - networks of power, 375
 - paradox of trust, 383–385
 - political influence, 376, 377
 - power of sovereign industries, 376
 - reality vs. real, 374
 - reflexivity, 374
 - risk as metaphysics, 381–383
 - sovereignty, 376
 - testimonies, 374
 - trauma, 374
 - world after, 387–388
- Fumaroles, 504
- G**
- Gamma distribution, 28, 657, 681
- Gamma modified Pareto distribution, 41
- Gas geochemistry, 508–509
- Gas slug, 471, 482, 483, 486, 489, 492, 493, 497
- Gaussian density, 656
- Gaussian distribution, 29, 37, 680
- Gaussian kernel, 671
- Generalized dimensions, 664
- General Systems Theory (GST), 720
- Geodesy, 504, 509–510, 539
- Geodetic leveling technique, 510
- Geological wedge, 56
- Geology, 509

Glacial isostatic adjustment (GIA), 539
 Global CMT, 10, 12, 344
 Global Navigation Satellite System (GNSS), 327, 328, 504, 539
 Global Positioning System (GPS), 504
 Global-scale tsunami hazard and risk assessment
 annual average loss (AAL), 235
 assumptions and uncertainties, 237–242
 earthquake source modeling, 218–222
 empirical run-up data use, 215–216
 inundation and exposed assets, 225–226
 mortality ratio, 226, 227
 offshore surface elevation, amplification of, 223–225
 overview, 216–218
 probabilistic loss estimation, 228–231
 probable maximum loss (PML), 235–237
 PTHA, 216, 217, 232
 scenario-based methods, 216
 tsunami propagation, 222–223
 vulnerability, 214, 226–228
 GPS Earth Observation Network (GEONET), 512
 Grassberger-Procaccia algorithm, 664
 2011 Great East Japan Earthquake, 341, 342, 346, 369
 Great subduction earthquake doublets, 274
 Kurils (2006, 2007), 275–276
 Samoa (2009), 277
 Green's function, 13, 82, 440, 458–459
 Green's law, 79–81, 352–355
 Green tensor, 474
 Grímsvötn volcano, 546–547

H

2012 Haida Gwaii earthquake, 86
 Hanging wall, 247
 Hazard curve, 213
 Heaviside function, 656
 Heavy minerals, 135
 Hekla, 547
 High-tech society model, 385–387
 Hillshade, 504
 Holocene volcanism, 649
 Homeland Security Alert Level, 733
 Hooke's law of elasticity, 58
 Horizontal lamination, 135
 Hough transform, 676
 Hubbert–Rubey fluid pressure ratio, 58
 Humanitarian Practice Network, 723
 Hurricane Katrina, 724
 Hurst exponent, 29, 32, 665
 Hurst number, 29
 Huygens' principle, 78
 Hybrid events, 425, 426
 Hydrodynamic forces, 187, 201, 207
 Hydrodynamic modelling, of tsunami evolution, 126–128
 Hydrostatic forces, 187, 202, 207
 Hyogo Framework for Action (HFA), 214
 Hypocenter, 247, 299, 337, 345
 depth and magnitude, 355

I

Iceland
 deformation, 543–545
 Grímsvötn fissure system, 541
 Hekla volcanic system, 542
 Katla volcanic system, 541
 volcano deformation, 545–555
 volcano geodesy, 542, 543
 Impulsive force, *see* Surge force
 2004 Indian Ocean tsunami, 72, 73, 76–78, 82, 87, 124, 726
 Inertial glut, 433–434
 Inference, 91
 Inflation, 504
 Initial Tsunami wavefield, 25
 Interferogram (radar), 505
 Interferometric synthetic-aperture radar (InSAR), 512–514, 539
 Intergovernmental Coordination Group/Pacific Tsunami Warning System (ICG/PTWS), 340
 International Decade for Natural Disaster Reduction (IDNDR), 726
 International Strategy for Disaster Reduction (ISDR), 726
 Inter-onset time, 656, 659, 664–666, 668, 669
 Interpolation method, 355
 Interseismic coupling, 248
 Inverse-Gaussian density, 658
 Inverse methods, 522
 Inversion/Inverse Problem, 247–248

J

Japan Meteorological Agency (JMA), 336, 339–347, 349, 353, 357–359, 361–368
 Japan, tsunami forecasting and warning, *see* Tsunami forecasting and warning
 1994 Java tsunami earthquake, 7
 Joint probability distribution, 91
 Jökulhlaup, 539

K

Katla volcano, 547
 Kernel density fusion cluster analysis, 677
 Kernel estimate, 671–672
 Kilauea and Mauna Loa volcanoes, deformation studies at, 529–530
 Kilauea Volcano, Hawaii, 709–710
 Kinematic earthquake model, 25
 Kinematic models, 25, 31, 47, 49
 Kinematic slip models, 31
 Knowledge-based society model, 385–387
 Krafla volcanic system, 549
 Krafla volcano, 554
 Kronecker's delta, 73
 Kronecker symbol, 428

L

Lagrangian approach, 130
 Laminae, 135
 Large eddy simulation (LES), 128

Late Mesolithic settlements, 153
 Lattice Boltzmann method (LBM), 496
 Lava-dome eruptions, 621–623, 626, 641
 Least square method, 344
 Leveling, 505
 instrument, 539
 Lévy α -stable distribution, 29
 Lidar, 505, 514
 Likelihood function, 91
 Linear time weakening friction, 32
 Linear wave theory, 122, 123
 Load-and-discharge model, 660
 Loading structures, 135
 Local magnitude (M_L) scale, 300
 Local tsunami, 187
 warning, 299
 Log-logistic density, 658
 Lognormal density, 657
 Lognormal distribution, 657
 Long-period (LP) event, 425, 426, 443, 449, 457, 458
 Long-period (LP) seismicity, volcanoes, 474–478
 inferred excitation mechanisms, 484–485
 magmatic-hydrothermal interactions, 478–481
 magmatic LP events, 481–484
 Long wave fission, 123
 Loss exceedance curve (LEC), 213, 230–231
 Loss return period, 213
 Love waves, 14
 Luminescence dating methods, 145

M

Mafic, 623
 Magma
 chamber, 539, 623, 624, 628–633, 635, 636, 638, 640–642
 and conduit resistance, 636–637
 discharge rate, 622, 623, 628, 631–633, 639, 640
 general steady-state solution, 630, 631
 intrusion, 712
 non-Newtonian properties, on eruption behaviour, 637–638
 semi-analytical approach, 629–630
 viscosity, 628, 629
 Magmatic-hydrothermal interactions, 478–481
 Magmatic LP events, 481–484
 Magma-water interactions (MWI), 573, 574
 Magnitude frequency distribution (MFD), 213
 Manning's coefficient, 127
 Mantle magnitude (M_m) method, 314
 Mantle-wave (M_m) magnitude, 299
 Marigram, 300
 Markov chain, 673–677
 Markov chain Monte Carlo (MCMC) methods, 96
 Markov processes, 659
 Maximum amplitude, of tsunami event, 26
 Maximum likelihood, 44
 Megathrust events, 248
 Bengkulu (2007), 259
 Maule (2010), 263–266

Nias (2005), 258–260
 Peru (2007), 261
 Santa Cruz Islands (2013), 270–274
 Sumatra-Andaman (2004), 256–258
 2004–2010 Sunda Trench sequence, 253–256
 Tohoku (2011), 266–270
 Melt, 622, 624, 629, 634–636
 Metaphysics, 381–383
 Method Of Splitting Tsunami (MOST), 126
 Micro-earthquakes (ME), 400, 419
 Microgravity surveys, 519–521
 Microlite, 631, 633, 634
 Microtextures, 135
 Mogi model, 522
 Molten fuel-coolant interactions (MFCIs), 572–574
 Moment magnitude, 248
 Moment-rate function, 430
 Moment tensor, 429–430, 472–474, 494
 cylindrical source, 432–433, 461–462
 slip on a fault, 430–431
 spherical source, 432, 459–461
 tensile crack, 431–432
 Momentum flux, 187
 Monte Carlo analysis, 96
 Monte Carlo simulation, 676, 692
 Monte Carlo-type procedure, 99
 Mortality rate, 213
 Mount Etna, 668–671
 Mount Pinatubo (Philippines), 706, 707
 Mount St. Helens (USA), 705, 706
 Mud drapes, 135
 Multifractal analysis, 664
 Multinomial distribution, 95

N

National Incident Management System (NIMS), 733, 735
 National Oceanic and Atmospheric Administration (NOAA), 340
 National Tsunami Hazard Mitigation Program, 126
 Navier-Stokes equations, 128, 627
 Nearest-neighbor kernel estimate, 672–673
 Nearshore tsunami evolution, 119–124
 Negative binomial distribution, 43
 Neoliberalism, 384
 Network-inversion filter, 524
 Newtonian behavior, 564
 Newtonian fluid, 629
 Newtonian liquid, 627, 637, 638
 Newtonian rheology, 637
 Newtonian viscosity, 564
 1992 Nicaragua earthquake, 6, 7, 11
 Non-dispersive waves, 122
 Non-homogeneous Poisson process, 654
 Nonlinear inversion methods, 83–84
 Nonlinear shallow-water (NSW) equations, 126
 Nonlinear shallow water (NLSW) models, 217
 Nonlinear tsunami, 122
 Non-Newtonian rheology, 633, 636
 Non-Newtonian viscosity, 564–567

Normal fault, 247
 Normal grading, 135
 Northwest Pacific Tsunami Advisory (NWPTA), 365
 Norwegian Sea, 155
 Notice to Airmen (NOTAM), 730
 Numerical deformation source models, 523–524

O

Omori-Utsu relation, 40
 Omori-Utsu temporal distribution, 43
 Onset time, 661, 666, 685
 Onshore tsunami deposits, sedimentological features of, 137–142
 Open conduit system, 661
 Optically stimulated luminescence dating, 145
 Ormen Lange gas field, 159, 160
 Ormen Lange project, 159
 Oscillating viscometer, 566
 Oscillatory seismicity, 625

P

Pacific Marine Environmental Laboratory (PMEL), 340
 Pacific Tsunami Warning Center (PTWC), 302
 observatory message, 304
 tsunami information statement, 304
 tsunami threat message, 304
 Palaeontology, 144–145
 Paleotsunami data, 44
 Palmieri electromagnetic seismograph, 394
 Parallel lamination, 135
 Parameter estimate, 654
 Pareto density, 655
 Pareto distribution, 28, 41–43, 45
 Particle size distribution, 585
 Peak ground displacement (PGD) measurements, 327
 Persistent scatterer InSAR (PS-InSAR), 514
 1996 Peru earthquake, 6
 Petrologic monitoring, 509
 Phase-field method, 472, 494
 Phenocryst, 633–636
 Philippine Institute of Volcanology and Seismology (PHIVOLCS), 706
 Photogrammetry, 517
 Phreatomagmatic basalt eruptions, 541
 Physical vulnerability, 213
 Plastic deformation, 505
 Plate Boundary Observatory (PBO), 512
 Plinian, 505, 507
 Plinian eruption, 601, 659
 Point process, 40, 42
 Poisson behavior, 653, 664
 Poisson distribution, 43, 653
Poisson-Gamma conjugate model, 95
Poisson-Gamma model, 95
 Poisson null hypothesis, 43
 Poisson process, 28, 44, 648, 653–654, 659
 Poroelasticity, 505
 Postseismic phase, 248
 Power-law density, 658

Power-law distribution, 33
 Power-law function, 32
 Power-law relation, 41
 Principal component analysis-based inversion method, 525
 Probabilistic analysis of tsunami hazards
 dynamic rupture models, 46–47
 framework of, 45
 stochastic source model implementation, 45
 Probabilistic tsunami hazard analysis (PTHA)
 distribution forms and mathematical techniques, 95–96
 epistemic and aleatory uncertainties, 96
 tsunami intensities, 92
 Probabilistic Tsunami Hazard Assessment (PTHA), 213, 214, 216–217, 232
 Probabilistic Tsunami Risk Assessment (PTRA), 213, 214
 Probability density function (PDF), 92, 213
 Probable maximum loss (PML), 213–214
 P-wave magnitude (pMag) scale, 300, 308
 P-wave moment magnitude (M_{wp}) method, 300, 309–314
 Pyroclastic flow, 416–418, 586–588, 608, 610, 621, 622, 633

Q

Quasi-periodic behavior, 44

R

Rabaul caldera (Papua New Guinea), 707–708
 Rabaul Volcanological Observatory (RVO), 707
 Radiocarbon ages, 153
 Rake, 247
 Random sea, 49
 Random wave, 27
 Random wavefield, 36
 Random wave theory, 28
 Rayleigh density distribution, 37
 Rayleigh distribution, 37
 Real-time tsunami forecasting model, 321–323
 Recharge, 505
 Recurrence interval/average return period, 92
 Regional tsunami warning, 300
 Renewal processes, 654–659
 Research and development (R&D) policies, 386–387
 Reverse Fault, 247
 Reykjanes Ridge (RR), 540
 Reynolds number, 636
 Rhyolite, 505
 Rift zones, 505, 539
 Rigidity/shear modulus, 248
 Rip-up clasts, 135
 RSAM records, 625, 626
 Rupture area, 248
 Rupture velocity, 248

S

Sanriku earthquake, 6
 Seafloor deformation, 73–74
 Sedimentological model, 165
 Sedimentology

- definition, 136
- sedimentary structures, 142–144 (*see also* Tsunami deposits)
- Sediment structures, 135
- Seismic body waves, 300
- Seismic coupling, 26
- Seismic cycle, 248
- Seismic data processing system, 345
- Seismic magnitude, 7
- Seismic methods, 306–307
- Seismic moment (M_0), 4, 8, 13, 25, 26, 248, 300
- Seismic signals
 - geological process, 395
 - source modeling, 398, 401–404
 - surface manifestations of volcanic activity, 404–410
 - volcanic eruption, 410–420
 - waveforms and spectra, 398–404
- Seismic sources, 436
 - inertial glut, 433–434
 - inversion for, 252–253
 - moment tensor, 429–433
 - single force, 434–437
 - stress glut, 427–428
 - and tsunamis, 252
- Seismic surface waves, 300
- Seismic waveforms, 345
- Seismic waves, 300
- Seismogeodetic data, 327
- Seismological analysis, 79
- Seismology, *see* Seismic signals
- Self-adjust range, 664
- Self affine, 29
 - fractals, 29
 - slip distribution, 32
- Self-similar 2-D function, 33
- Self-similar fractal, 29
- Self-similarity, 662
- Self-similar random function, 32
- Semicontinuous GPS (SGPS), 511
- Semipermanent GPS (SPGPS), 511
- Shallow earthquake, 300
- Shallow-water equation models, 127
- Shallow water theory, 75
- Shallow-water-wave theory, 187
- Shiveluch volcano, 621, 638
- Shoaling effect, 187
- Shoreline algorithms, 128–131
- Short Message Service (SMS), 730
- Short-term average/long-term average (STA/LTA), 343
- Silicic, 636
- Sills, 505
- Simple Poisson process, 653
- Simulation point, 350, 351, 355–357
- Size-predictable model, 661
- Slip distribution, 248
- Slip-weakening distance, 32
- Small baseline subset InSAR (SBAS), 514
- Solitary wave fission, 123
- Solomon Island earthquake, 6
- Sompi method, 443, 447
- Soufrière Hills Volcano (SHV), 638–640
- Source potency, 26
- Source-time function, 430
- Spatial aspects, volcanic eruptions
 - alignments and clusters, 676–678
 - kernel estimate, 671–672
 - Markov chain, 673–677
 - nearest-neighbor kernel estimate, 672–673
 - spatiotemporal intensities, 670–673
 - spatiotemporal nearest-neighbor estimate, 670, 671
- Spatiotemporal nearest-neighbor estimate, 670, 671
- Specific mechanical energy (SME), 586, 587
- Spectral analysis, 442–449
- Splay fault, 32, 47
 - segments, 32
 - systems, 31
- Static earthquake model, 26
- Static slip models, 28–30
- Static surface displacements, 30
- Static Tsunami generation model, 26
- Still, 539
- Stochastic branching processes, 28
- Stochastic initial stress distribution, 33
- Stochastic prestress, 47
- Stochastic shear stress distributions, 34
- Stochastic slip model, 28, 29, 45, 49
- Stochastic stress distributions, 33
- Stochastic system, 664
- Storegga tsunami, 154
 - coastal lakes in Norway, 163–168, 171
 - definition, 154
 - deposits, 155
 - estuaries in Scotland, 168–172
 - Mesolithic, 153
 - numerical simulation, 175–178
 - peat outcrops on Shetland, 172–175
 - radiocarbon ages, 153
 - retrogressive slide motion, 153
 - run-up, 153
 - sediment trap, 153
 - slide events, 155
 - slide information, 161
 - slides, 158–162
 - Stone Age humans, 181–183
 - tsunami deposits and run-up estimation, 156
 - tsunami events, 178–181
- Storm deposits, 137
- Strain, 248
- Strainmeter(s), 506, 518, 519
- Stress, 248
- Stress drop, 29, 32, 33
 - pattern, 31
- Stress glut, 427–428
- Strike, 247
- Strike-slip events, Sumatra (2012), 278–280
- Strike-slip fault, 247

- Strombolian eruptions, 596, 602, 604, 606
 Subduction earthquake, 62–64
 Subduction zone, 26, 187, 248, 251
 environments, 248
 Sullom Voe, 175
 2004 Sumatra-Andaman earthquake, 74, 78, 79, 81, 84
 2004 Sumatra-Andaman Islands, 4
 Sumatra-Andaman 2004 megathrust earthquake, 250
 Sum of Asymptotic Mean Squared Error (SAMSE), 672, 681, 691
 Supplemental tsunami threat messages, 304
 Surface-wave magnitude (M_S), 299
 Surge force (impulsive force), 187
 Surge forces, 204, 207
 Systematic geoscience studies, 701
 System of equations, 633–636
- T**
- Technological democracy, 383
 Teleseismic windowing scheme, 325
 Teletsunami warning, 300
 Temporal clustering, 28, 43
 Temporal models, volcanic eruptions, 650–653
 chaos and fractals, 662–667
 Markov processes, 659
 Poisson processes, 653–654
 renewal processes, 654–659
 time-and size-predictable models, 659–662
 Tensile crack, 431–432
 Terzaghi's equation, 203
 Thermodynamic properties, 496
 Thermoelasticity, 506
 Three Sisters volcanic center, Oregon, 525–527
 Thrust fault, 247
 Tide gauge, 26
 Tiltmeters, 506, 518
 Time-predictable behavior, 659
 Time series, 37, 49
 Time-series analysis, 651
 2011 Tohoku earthquake, 4, 6, 10
 2011 Tohoku earthquake tsunami, 83
 Tohoku event, 250
 2011 Tohoku-Oki earthquake, 30
 2011 Tohoku tsunami, 42
 1968 Tokachi-oki earthquake, 79–82
 Transpressive Fault, 247
 Trench, 248
 Triangulation, 506, 510
 Trilateration, 506, 510
 Trust, 383–385
 Tsunami(s), 72, 300
 amplitude, 26, 248, 340–342, 346, 347, 349, 352, 353, 355–357, 363, 366, 368
 arrival time, 26
 catalog data, 28
 catalogs, 26, 27, 41, 42, 49
 characteristics, 192–199
 coda, 26, 36–38, 40
 complexities, 27
 definition, 187–188, 301
 dynamic generation models, 27
 dynamic rupture models, 27
 earthquake, 27, 300, 303, 317–321
 earthquake fault parameters, estimation of, 81–86
 elastic deformation models, 27
 event catalog, 44
 forces, prediction of, 201–206
 generation, 73–75
 generation, propagation, and run-up modeling, 199–200
 geometric spreading and dispersion, 27
 Green's law and tsunami heights, 79–81
 hazard prediction, 206
 height, 26
 instrumental data, 76
 inverse refraction diagram, 78–80
 laboratory-derived frictional parameterizations, 27
 magnitude, 81
 modern, historical and prehistoric tsunami heights, 76–78
 monitoring system, 363–365
 numerical computations, 75–76
 origin time, 26
 Pareto distribution, 28
 power-law scaling, 28
 propagation, 75, 118, 124, 126, 128
 PTHA, 45–47
 random wave and vibration theory, 27
 random wavefield, 28
 refraction diagram, 78–79
 risk, 214
 runup, 26
 scattering and reflection, 27
 and seismic source, 252
 shallow-water nonlinear interactions, 121
 short-period energy, 121
 static and dynamic rupture models, 28
 static generation models, 27
 static tsunami generation models, 27
 stochastic models, 36–45
 stochastic processes, 27
 time series, 28, 36–40
 time series observations, 28
 warning centers, 301
 wavelength, 249
 waves, 301
- Tsunami deposits, 135
 age estimation methods, 145–146
 characterization, 136
 description, 137
 grain size characteristics, 140
 identification, sedimentological criteria for, 140
 nature of, 136
 onshore, 137–142
 palaeocurrent indicators, 143
 preservation of, 145
 reexamination studies, 147

- Tsunami Early Warning System (TEWS), 339–342
- Tsunami earthquakes, 4, 5, 12, 18–19, 280
 - aftershocks, 10–12
 - horizontal deformation of ocean floor, 15, 16
 - hypocenters, 10
 - Java (2006), 282
 - low shear modulus, weak materials with, 13–14
 - Mentawai (2010), 282–284
 - presence of fluids, 17–18
 - shallow depth of slip, 14
 - shallow fault dip, 14–16
 - slip/fault models, 10, 11
 - slow character, 7–10, 13
 - subduction of bathymetric features, 16–17
 - uplift, slides/splay faulting, 17
- Tsunami effects, in man-made environment
 - damage in coastal areas, 190–192
 - evacuation strategy, 188, 208
 - tsunami-damage mitigation strategy, 188–189
 - tsunami-resistant coastal structures, 189–190
- Tsunami forecasting and warning
 - CMT solutions, 367–368
 - coastal block partitioning for tsunami forecast, in Japan, 347, 348
 - complicated bathymetry, 339
 - derivation of tsunami amplitude, 352–353
 - distant event, 364, 365
 - EEW, 366, 367
 - fault dip, strike and slip angle, 339
 - fault length, width and average slip amount, 338
 - fault parameter setting, 350–351
 - Green's law application, adequacy of, 353–356
 - hypocenter depth and magnitude, 355
 - hypocentral location, 349–350
 - initial tsunami wave distribution, uncertainty of, 337–338
 - interpolation method, 355
 - magnitude underestimation, 342
 - maximum risk method, 355–357
 - numerical simulation of tsunami propagation, 351–352
 - NWPTA, 365
 - prompt Mw calculation failure and insufficient warning, 342
 - real-time seismic data processing system, 343–346
 - seismic network, 342–343
 - tsunami arrival time estimation, 357–366
 - tsunami database creation, 349
 - tsunami forecast assembling, 357–363
 - tsunami forecast criterion and category, 347, 349
 - tsunami forecast dissemination, 363
 - tsunami information statement, 342
 - tsunami monitoring system, 363–365
 - warning statement, 342
- Tsunami Forecasting based on Inversion for initial sea-Surface Height (tFISH), 84
- Tsunamiogenic coseismic seafloor deformation, 64–66
- Tsunamiogenic major and great earthquakes
 - great subduction earthquake doublets, 274–277
 - Haida Gwaii (2012), 287–290
 - megathrust events, 253–274
 - Solomon Islands (2007), 284–287
 - strike-slip events, 278–280
- Tsunami height, 249
- Tsunami inundation
 - bathymetric and topographical features, 124–126
 - depth and velocity, 199
 - hydrodynamic modeling of tsunami evolution, 126–128
 - moving shoreline algorithms, 128–131
 - nearshore tsunami evolution, physics of, 119–124
 - tsunami generation and open ocean propagation, 118–119
- Tsunami warning system, 300
 - challenge, 303
 - components, 303
- TUNAMI-N2, 126
- Two-point azimuth method, 676
- U**
 - Udden–Wentworth scale, 137
 - UN ISDR Platform for the Promotion of Early Warning (PPEW), 726
 - United States Geological Survey (USGS), 717, 728, 734–736
 - US Global Positioning System (GPS), 539
 - U.S. National Volcano Early Warning System (NVEWS), 728
- V**
 - Value-added society model, 385–387
 - Velocity-weakening behavior, 60, 61
 - Vertical evacuation structure, 187
 - Very-long-period (VLP) event, 425–427, 443, 449, 457, 458
 - Very-long-period (VLP) signals, volcanoes, 400, 485–486
 - coupled diffusive-elastic pressurization, at Popocatepetl Volcano, 486–489
 - slug disruption, at Stromboli, 489–493
 - Viscoelasticity, 506, 567
 - Volatile, 624, 627, 634, 641
 - diffusion, 570
 - solubility, 568, 569
 - Volcanic Alert, 413
 - Volcanic Ash Advisory (VAA), 730
 - Volcanic Ash Advisory Centres (VAAC), 735
 - Volcanic crisis, 702–704, 706–708, 711
 - Volcanic eruptions, 393, 410–420, 648
 - data, 649–650
 - forecasting, 648
 - interactions with earthquakes, 682–689
 - model assessment, 687, 690–692
 - spatial aspects, 670–677
 - temporal models, 650–666
 - volcanic regimes, 666–671
 - Yucca Mountain, 677, 679–682
 - Volcanic eruptions, explosive
 - caldera collapse, 610–612
 - cinder cones, 607, 609

- Volcanic eruptions, explosive (*cont.*)
 - consequences, 571
 - decompression-driven flows, 576, 578–581, 590–596
 - degassing-driven flows, 594, 596–602
 - explosion consequences, 606–612
 - exsolution, 574–577
 - fall velocity of silicate fragments, 583
 - fragment electrification, 583
 - fragment size distributions, 581–583
 - high VEI events, 588–589
 - hydrovolcanism, 572–574
 - low VEI events, 602–607
 - magma rheology, 563–568
 - nucleation and diffusion, 568–570
 - permeability, 570–571
 - post-fragmentation flows, 589–590
 - pyroclastic flows, 586–588, 608, 610
 - silicate fragments and aggregation, 583–588
 - volatile solubility, 568, 569
- Volcanic explosions, 418–419
- Volcanic Explosivity Index (VEI), 563, 613, 649, 652, 661, 662, 666, 686, 687
 - high VEI events, 588–589
 - low VEI events, 602–607
- Volcanic hazards and early warning
 - dilemma of “false alarms,” 711
 - emergency-management authorities, 702, 703
 - geologic mapping and dating studies of volcanoes, 710
 - outcomes of volcano unrest, 702–710
 - predictive capability, 710–711
 - volcano monitoring, 710–711
- Volcanic hazards warnings
 - analytical tools, 729
 - diversity of, 717–718
 - early warning systems (EWS), 719–728
 - effectiveness of, 731–733
 - reconceptualising, 737–738
 - social contexts, 738
 - standardisation, 733–736
 - VALS, 729–731
 - volcano mitigation and communication strategies, 718–719
 - warning tools, 729
- Volcanic materials
 - magma rheology, 563–568
 - Newtonian viscosity, 564
 - non-Newtonian viscosity, 564–567
 - nucleation and diffusion, 568–570
 - permeability, 570–571
 - viscosity/modulus, 566–568
 - volatile solubility, 568, 569
- Volcanic processes, 571–572
 - explosion products, 581–583
 - explosive processes, 572–581
 - silicate fragments and aggregation, 583–588
- Volcanic regimes, 666–671
- Volcanic seismology, 394
- Volcanic system, 539
- Volcanic tremor, 400, 416
- Volcano, 393
 - monitoring, 701–703, 705–711
- Volcano Alert Level System (VALS)
 - aviation colour code, 729
 - event driven (urgent) messages, 730
 - information statements, 730–731
 - time driven (scheduled) status messages, 730
- Volcano Alert Notification' (VAN), 730
- Volcano Deformation in Iceland
 - Bárdarbunga volcano, 551–552
 - Eyjafjallajökull, 552
 - Grímsvötn volcano, 546–547
 - Hekla, 547
 - Katla volcano, 547
 - Krafla volcanic system, 549
- Volcano deformation measuring techniques
 - geodetic leveling, 510
 - Global Positioning System, 510
 - InSAR, 512–514
 - lidar, 514–516
 - microgravity surveys, 519–521
 - photogrammetry, 516–517
 - strainmeters, 518, 519
 - tiltmeters, 518
 - triangulation, 510
 - trilateration, 510
- Volcano-deformation modeling, 521
- Volcanoes, 506
 - LP seismicity, sources of, 474–485
 - seismic sources, 472–474
 - VLP signals, source processes of, 485–493
- Volcano geodesy, 506
 - in Iceland, 542
 - inverse problem in, 524–525
- Volcanology, 507
- Volcano observatories, 507, 508
- Volcano Observatory Notice for Aviation' (VONA), 730
- Volcano-seismic signals, 425, 426, 437, 440, 442, 449, 458
- Volcano seismology, 508
- Volcano-tectonic (VT) earthquake, 425, 426, 447
- Volcano unrest
 - culmination in major eruption/return to dormancy, long lull in unrest, 703–704
 - culmination in major eruption/return to dormancy, short duration of unrest, 702, 703
 - Kilauea Volcano, Hawaii, 709–710
 - Mount Pinatubo (Philippines), 706, 707
 - Mount St. Helens (USA), 705, 706
 - ongoing irregular long-duration volcano unrest, 708, 709
 - ongoing unrest, long periods, 704, 705
 - Rabaul caldera (Papua New Guinea), 707–708
- Volume of fluid (VOF) method, 128
- Vulnerability, 214, 226–228

W

Waveform inversion, 82–85, 437–442, 473, 474

Wedge, 248

Wedge mechanics

dynamic Coulomb wedge, 60–62

stable and critical Coulomb wedges, 58–60

subduction earthquake, stress drop and increase in,
62–64

Weibull density, 656

Weibull distribution, 656, 657

Weibull process, 654, 680

World Conference on Natural Disaster Reduction, 723

World Organization of Volcano Observatories
(WOVO), 712

W-phase, 301

Y

Yellowstone magmatic system, Wyoming, 527–529

Yucca Mountain, 677, 679–682